FeTiMM – A new oxybarometer for mafic to felsic magmas

R. Arató**, A. Audétat**

Abstract

The oxidation state of magmas is a key parameter that is notoriously difficult to reconstruct. The most common approach is via magnetite–ilmenite oxybarometry. However, many natural magmas do not contain ilmenite, preventing application of this technique. Here we present a new method that allows \( f_O^2 \) to be reconstructed based on the partitioning of Fe and Ti between magnetite and silicate melt. The new method, which we call FeTiMM, is applicable to both ilmenite-free and ilmenite-bearing samples, and even to slowly cooled intrusive rocks such as granites. FeTiMM was calibrated on 109 experiments covering a wide range of oxygen fugacities, temperatures, pressures and silicate melts ranging from basaltic to rhyolitic composition, and returned \( f_O^2 \) values that agree within 0.5 log units with independently constrained \( f_O^2 \) values in all but five cases. A first test on 19 natural samples of dacitic to rhyolitic compositions was equally successful. FeTiMM thus opens the door for numerous new applications in various disciplines of Earth Sciences, including the fields of volcanology, igneous petrology, experimental geochemistry and ore geology.

Introduction

Oxygen fugacity is an important thermodynamic parameter in magmatic systems because it exerts a first order control on phase equilibria as well as on mineral–melt and fluid–melt partition coefficients. The most commonly used and most reliable way to reconstruct magmatic \( f_O^2 \) is via magnetite–ilmenite oxybarometry (e.g., Buddington and Lindsley, 1964; Carmichael, 1967; Stormer, 1983; Andersen and Lindsley, 1988; Ghiorso and Sack, 1991; Lattard et al., 2005; Ghiorso and Evans, 2008). Alternative approaches are based on (i) the \( \text{Fe}^{2+}/\text{Fe}^{3+} \) ratio of whole rocks (Kress and Carmichael, 1989; Putirka, 2016), (ii) mineral reactions involving olivine, pyroxene, and/or sphene (Frost and Lindsley, 1992; Lindsley and Frost, 1992; Andersen et al., 1993; Xirouchakis et al., 2001), (iii) biotite, K-feldspar and magnetite (Wones and Eugster, 1965; Wones, 1981), (iv) zircon Ce (and Eu) anomalies (Ballard et al., 2002; Trail et al., 2012; Smythe and Brenan, 2016), and (v) single amphibole oxybarometry (Ridolfi et al., 2010). However, despite these various approaches, reconstruction of magmatic \( f_O^2 \) in igneous rocks remains difficult, particularly in the case of intrusives, because during slow cooling Fe-Ti-oxide minerals usually get either reset or altered. Furthermore, many magmas do not contain ilmenite but only magnetite (Frost and Lindsley, 1991), precluding the application of magnetite–ilmenite oxybarometry.

Calibration of the FeTiMM Oxybarometer

The aim of this study was to develop an oxybarometer that is based on element partitioning between a mineral and silicate melt, such that it can be applied to samples in which these phases occur as inclusions within phenocrysts and thus were protected from re-equilibration and alteration during slow cooling. Iron partitioning between magnetite and melt is a promising candidate because magnetite is a common mineral and because magnetite solubility in silicic melts has been shown to depend on \( f_O^2 \) (Gaillard et al., 2001). However, magnetite solubility (and thus Fe partitioning between magnetite and silicate melt) also depends strongly on temperature \( (T) \) and melt composition. At constant \( f_O^2 \) and \( T \), magnetite solubility increases by a factor of up to 6 as the alumina saturation index (ASI; defined as the molar Al\(_2\)O\(_3\)/(Na\(_2\)O + K\(_2\)O + CaO) ratio) decreases from 1.0 to 0.6. Magnetite solubility is thus not suitable as an oxybarometer unless \( T \) and ASI can be extremely well constrained, which is commonly difficult in natural samples (Arató and Audétat, 2017a). However, we noticed that the effect of ASI on magnetite solubility is similar to that on TiO\(_2\) solubility (Kularatne and Audétat, 2014), with the latter being independent on \( f_O^2 \). Hence, the effect of melt composition may be diminished by dividing the mgt–melt partition coefficient of Fe (\( D_{\text{mgt/melt}} \)) by that of Ti (\( D_{\text{mgt/melt}} \)).

We tested this idea first on a set of 50 of our own experiments conducted in the system magnetite–H\(_2\)O–rhyolite melt at various oxygen fugacities, temperatures, pressures, melt ASIs, and magnetite compositions (Arató and Audétat, 2017a). However, despite these various approaches, reconstruction of magmatic \( f_O^2 \) in igneous rocks remains difficult, particularly in the case of intrusives, because during slow cooling Fe-Ti-oxide minerals usually get either reset or altered. Furthermore, many magmas do not contain ilmenite but only magnetite (Frost and Lindsley, 1991), precluding the application of magnetite–ilmenite oxybarometry.

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The results (Fig. S-1) revealed that the melt composition effect indeed gets greatly reduced in this way, and that the Fe–Ti exchange coefficient between magnetite and silicate melt, \(D_{\text{Fe-Ti}}^{\text{melt/melt}} = (D_{\text{FeO}}^{\text{melt}})/(D_{\text{TiO}_2}^{\text{melt}})\), depends most strongly on \(f_\text{O}_2\) with the effect of temperature becoming negligible. In a second step we extended the dataset by 59 experiments from 14 different studies performed at 750-1100 °C, 0.1-700 MPa, oxygen fugacities of -1.3 to +5.5 log units relative to the fayalite-magnetite-quartz buffer (\(\Delta FMQ-1.3 \) to \(\Delta FMQ+5.5\)), with melt compositions of 48-79 wt. % \(\text{SiO}_2\) and ASI = 0.3-1.3, and magnetite compositions of 0.01-28 wt. % \(\text{TiO}_2\). These 59 experiments were left after screening 35 experimental studies with a total of >1600 experiments (Table S-2) for the following criteria: (i) \(f_\text{O}_2\) was controlled experimentally, (ii) magnetite coexists with ilmenite in at least some of the experiments, (iii) compositional data for magnetite, silicate melt, and ilmenite are available, (iv) the reported average compositions of magnetite and ilmenite represent equilibrium pairs (Bacon and Hirschmann, 1988), and (v) the average compositions of magnetite and silicate melt pass a similar test that we developed to check for magnetite–melt equilibrium (Fig. S-2). In this second step, we focused on ilmenite-saturated experiments to be able to constrain \(f_\text{O}_2\) values independently via magnetite–ilmenite oxybarometry. In nine cases, the reported experimental \(f_\text{O}_2\) values deviated by more than 1.0 log unit \(f_\text{O}_2\) from the values obtained via magnetite–ilmenite oxybarometry (Ghiorso and Evans, 2008; Fig. S-3), suggesting problems with the control of experimental \(f_\text{O}_2\) (Matjuschkin et al., 2015). We thus relied on the \(f_\text{O}_2\) values calculated via magnetite–ilmenite oxybarometry for all 59 ilmenite-saturated experiments. To account for the large range of melt compositions, it was necessary to include MgO in the melt compositional parameter. Based on the extended dataset of 109 experiments we developed a model (which we hereinafter call FeTiMM) that allows \(f_\text{O}_2\) to be calculated as a function of \(D_{\text{Fe-Ti}}^{\text{melt/melt}}\) (with \(f_\text{O}_2\) and \(\text{TiO}_2\) measured in weight percent, and the melt composition reported dry) and the melt compositional parameter AMCNK = molar \(\text{Al}_2\text{O}_3/(\text{CaO + Na}_2\text{O + K}_2\text{O})\):

\[
\Delta FMQ = (\log(D_{\text{FeO}}^{\text{melt}})/(D_{\text{TiO}_2}^{\text{melt}})) + 0.137*\text{AMCNK} + 0.102)/(0.288*\text{AMCNK} + 0.054)
\]

The rationale behind this equation is depicted in Figure 1. The overall uncertainty of the FeTiMM model, calculated from the errors of the fits in Figure 1b,c propagated into Equation 1 (see Supplementary Information) increases from \(\pm 0.2\) to \(\pm 0.5\) log units \(f_\text{O}_2\) at \(\Delta FMQ +1.5\), to \(\pm 0.3\) to \(\pm 0.5\) log units \(f_\text{O}_2\) at \(\Delta FMQ +4.5\) (Supplementary Table S-1). The performance of FeTiMM on the 59 ilmenite-saturated experiments is shown in Figure 2. As explained above, we relied on \(f_\text{O}_2\) values calculated via magnetite–ilmenite oxybarometry (Ghiorso and Evans, 2008) for this test. The uncertainty of the latter model was not explicitly stated (Ghiorso and Evans, 2008), but based on their Figure 27 it can be estimated at \(\pm 0.3\) log units at reducing conditions (\(\Delta FMQ -1\) to \(\Delta FMQ +1\)) to \(\pm 0.5\) log units at strongly oxidising conditions (\(\Delta FMQ +4.5\); not considering a group of outliers), i.e. similar to the uncertainty associated with our model. Within these errors, 63 out of the 109 experiments show perfect agreement between the two methods and all but five experiments return \(f_\text{O}_2\) values that agree within \(\pm 0.5\) log units. No correlations are evident between the degree of correspondence and \(f_\text{O}_2\), temperature, melt \(\text{SiO}_2\) content, ASI, or magnetite composition (Fig. S-4), suggesting that FeTiMM works equally well over the entire range of the investigated P-T-X conditions. It should be mentioned that any misfit can result from various sources including (i) analytical errors, (ii) experimental problems, as well as (iii) weaknesses in either model.

![Figure 1](image-url) Development of the FeTiMM model. (a) 109 experimental data points (59 ilmenite-saturated; 50 ilmenite-undersaturated) were split into five groups of similar oxygen fugacity, through which linear regressions were fit. \(D_{\text{Fe-Ti}}^{\text{melt/melt}} = (D_{\text{FeO}}^{\text{melt}})/(D_{\text{TiO}_2}^{\text{melt}})\). AMCNK = molar \(\text{Al}_2\text{O}_3/(\text{CaO + Na}_2\text{O + K}_2\text{O})\). (b) Variation of the slopes and intercepts of the linear fits in (a) as a function of oxygen fugacity expressed in log unit deviations from the fayalite-magnetite-quartz (FMQ) buffer.

Tests on Additional Ilmenite-Undersaturated Experiments

As discussed above, Equation 1 was derived from a database that comprises 50 ilmenite-undersaturated, rhyolitic experiments, plus 59 ilmenite-saturated, basaltic to rhyolitic experiments. It remains to be tested whether FeTiMM works also for ilmenite-undersaturated intermediate to mafic magmas. However, in ilmenite-undersaturated experiments $f_{O_2}$ cannot be independently constrained via magnetite–ilmenite thermobarometry, which forces us to rely on reported experimental $f_{O_2}$ values even if the test with the ilmenite-undersaturated experiments (Fig. S-3) revealed that these values are not always reliable. To reduce the chance of including erroneous $f_{O_2}$ reference values, we restricted our choice of ilmenite-undersaturated experiments to studies which comprised both ilmenite-undersaturated and ilmenite-saturated experiments, and which in the latter case showed good agreement between the reported experimental and magnetite–ilmenite-based $f_{O_2}$ values. This approach returned 27 data points from 5 different studies (Table S-1), excluding 22 data points that did not pass the Mn/Mg magnetite–melt equilibrium test mentioned above. The performance of FeTiMM on these 27 literature-based experiments plus 50 of our own ilmenite-undersaturated experiments is shown in Figure 3. The data in Figure 3 show a larger scatter than those in Figure 2, which is likely due to errors in the reported experimental $f_{O_2}$ values (cf. Fig. S-3). Nevertheless, within the error quoted for FeTiMM, 7 of the 27 experiments return $f_{O_2}$ values that agree between the two methods and all but 5 experiments show a correspondence within 0.5 log units. Again, no correlations are evident between the degree of $f_{O_2}$ correspondence and other key variables (Fig. S-5), suggesting that FeTiMM works equally well for mafic as for silicic rocks.

Application to Natural Samples

The results of a first application of FeTiMM to 19 natural samples of rhyolitic to dacitic composition are shown in Figure 4. Details on the samples and analytical methods can be found in Arató and Audétat (2017a). All samples were ilmenite-saturated, such that $f_{O_2}$ could be independently constrained via magnetite–ilmenite oxybarometry. Coexistence of analysed magnetite, ilmenite and silicate melt was verified by means of the Mg/Mn magnetite–ilmenite partitioning test (Bacon and Hirschmann, 1988) plus our own Mn/Mg magnetite–melt partitioning test described in the Supplementary Information. In all but one sample, FeTiMM returned $f_{O_2}$ values that agree within 0.5 log units with those obtained via magnetite–ilmenite oxybarometry (Fig. 4).

One of the main advantages of FeTiMM is that it can be applied to magmas that do not contain ilmenite, which is true for many igneous rocks of mafic to felsic composition, particularly for those that are alkali-rich (Lindsley and Frost, 1992). Another major advantage of the method is that it can be applied to slowly cooled and/or altered rocks if magnetite and silicate melt are present as inclusions within phenocrysts (preferably quartz) and are analysed as entities by LA-ICP-MS, thereby effectively reversing compositional heterogeneities.
that developed within the inclusions during slow cooling. The sole disadvantage of FeTiMM is that it requires knowledge of the silicate melt composition. This can be readily accomplished in rapidly quenched, volcanic samples with glassy matrix, but is a bit more difficult in holocrystalline, porphyritic samples (for fresh samples selective analysis of the matrix suffices), and can be challenging in holocrystalline, coarse grained samples. In the latter samples, both magnetite and silicate melt need to be analysed as inclusions within phenocrysts, with the quantification of melt compositions requiring re-homogenisation experiments and/or constraints from whole rock data (Audétad and Lowenstern, 2014; Arató and Audétat, 2017a).

Figure 4 Application of FeTiMM to a set of 19 natural, ilmenite-saturated samples of rhyolitic to dacitic composition (Arató and Audétat, 2017a). Oxygen fugacities (reported in log units relative to the fayalite–magnetite–quartz buffer) obtained via FeTiMM agree within 0.5 log units with those obtained via magnetite–ilmenite oxybarometry (Ghiorso and Evans, 2008) in all but one case. Error bars denote 1σ standard deviation of the fO2 values obtained from several magnetite–melt pairs.

References


Lattard, D., Sauzeaud, U., Kasemann, M. (2005) New calibration data for the Fe-Ti oxide thermo-oxybarometers from experiments in the Fe-Ti-O system at 1 bar, 1,000–1,300°C and a large range of oxygen fugacities. Contributions to Mineralogy and Petrology 149, 735–754.


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**Supplementary Information**

The Supplementary Information includes:

- Mn/Mg Test to Check for Equilibrium between Magnetite and Silicate Melt
- Estimation of the Uncertainty of FeTiMM
- Figures S-1 to S-5
- Tables S-1 and S-2
- Supplementary Information References

**Mn/Mg Test to Check for Equilibrium between Magnetite and Silicate Melt**

To be able to check for equilibrium between a given magnetite–silicate melt pair, a new test was developed which is based on the same two elements (Mn, Mg) as the equilibrium test between magnetite and ilmenite (Bacon and Hirschmann, 1988), although the underlying rationale is quite different. We searched for two elements whose magnetite–melt partition coefficients depend in a similar manner on melt composition, but whose absolute concentrations in the silicate melt vary strongly during magma differentiation. With this approach, melts that do not coexist with a given magnetite composition should be easily identifiable. We thus use the Mn–Mg exchange coefficient $D_{\text{Mn–Mg,melt}} = (D_{\text{MnO,melt}})/(D_{\text{MgO,melt}})$, with MnO and MgO given in weight percent. A plot of $D_{\text{Mn–Mg,melt}}$ versus MnO/MgO ratio in the silicate melt for 356 magnetite–melt pairs with melts ranging from basaltic composition (48 wt. % SiO$_2$) to high silica rhyolitic composition (79 wt. % SiO$_2$) is shown in Figure S-2. It turns out that 95 % of the data points with reported Mn and Mg concentrations display a $D_{\text{Mn–Mg,melt}}$ value between 1 and 4, with the remaining 5 % apparently representing outliers because no correlation between the magnitude of mismatch and the MnO/MgO ratio in the silicate melt (Fig. S-2) or other parameters (Table S-1) is evident.

**Estimation of the Uncertainty of FeTiMM**

We assume that $f$ is an $n$-dimensional differentiable function and $X_1, X_2, ..., X_n$ are independent variables with standard deviations $\sigma_1, \sigma_2, ..., \sigma_n$. Therefore, the overall uncertainty can be estimated from the first order linear estimate of $f(X_1, X_2, ..., X_n)$:

$$
\sigma_{x_1, x_2, ..., x_n} = \sqrt{\sum_{i=1}^{n} \left( \frac{\partial f}{\partial X_i} (EX_1, EX_2, ..., EX_n) \right)^2 \sigma_i^2}, \quad \text{Eq. S-1}
$$

where $\frac{\partial f}{\partial X_i}$ is the $i$-th partial derivative of the function $f$.

If the expected values (EX) of the variables (in our case chemical analyses) are unknown, one can replace the expected values by the variables themselves:

$$
D(f(X_1, X_2, ..., X_n)) \approx \sqrt{\sum_{i=1}^{n} \left( \frac{\partial f}{\partial X_i} (X_1, X_2, ..., X_n) \right)^2 \sigma_i^2}. \quad \text{Eq. S-2}
$$

Our function has the following form:

$$
\Delta \text{FMQ} = \log_{10}\left( \frac{\text{FeO}_{\text{tot,melt}}}{\text{FeO}_{\text{tot,melt}}} \frac{\text{TiO}_2}{\text{TiO}_2} \frac{\text{Al}_2\text{O}_3, \text{Denom}}{a \cdot \text{AMCNK} + c} \right),
$$

where

$$
\frac{\text{FeO}_{\text{tot,melt}}}{\text{FeO}_{\text{tot,melt}}} = \frac{\text{FeO}_{\text{tot,melt}}}{\text{FeO}_{\text{tot,melt}}} \frac{\text{TiO}_2}{\text{TiO}_2} \frac{\text{Al}_2\text{O}_3, \text{Denom}}{a \cdot \text{AMCNK} + c}
$$

and

$$
\text{AMCNK} = \frac{\text{Al}_2\text{O}_3, \text{Denom}}{\text{Denom}}.
$$

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The variables CaO, Na$_2$O, K$_2$O, and MgO appear only as a sum in the equation (i.e. as CaO + Na$_2$O + K$_2$O + MgO) and their variances are independent, therefore they can be treated as a single variable in the calculation of the partial derivatives – named “Denom” – the variance of which is the sum of the individual elements variances:

\[
\text{Denom} = \text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O} + \text{MgO},
\]

\[
\sigma^2_{\text{Denom}} = \sigma^2_{\text{CaO}} + \sigma^2_{\text{Na}_2\text{O}} + \sigma^2_{\text{K}_2\text{O}} + \sigma^2_{\text{MgO}}.
\]

It is important to note that Al$_2$O$_3$, CaO, Na$_2$O, K$_2$O, and MgO (as well as their standard deviations) refer to molar abundances, whereas FeO$_{\text{tot mgt}}$, FeO$_{\text{tot melt}}$, TiO$_2$$_{\text{mgt}}$ and TiO$_2$$_{\text{melt}}$ are given in weight percent.

The partial derivatives used for the propagation of uncertainty are the following:

a) For the analytical uncertainty:

\[
\frac{\partial \Delta \text{FMQ}}{\partial \text{FeO}_{\text{tot mgt}}} = \frac{1}{\text{FeO}_{\text{tot mgt}} \cdot \ln(10) \cdot (a \cdot \text{AMCNK} + c)}
\]

\[
\frac{\partial \Delta \text{FMQ}}{\partial \text{FeO}_{\text{tot melt}}} = \frac{-1}{\text{FeO}_{\text{tot melt}} \cdot \ln(10) \cdot (a \cdot \text{AMCNK} + c)}
\]

\[
\frac{\partial \Delta \text{FMQ}}{\partial \text{TiO}_2_{\text{mgt}}} = \frac{1}{\text{TiO}_2_{\text{mgt}} \cdot \ln(10) \cdot (a \cdot \text{AMCNK} + c)}
\]

\[
\frac{\partial \Delta \text{FMQ}}{\partial \text{TiO}_2_{\text{melt}}} = \frac{-1}{\text{TiO}_2_{\text{melt}} \cdot \ln(10) \cdot (a \cdot \text{AMCNK} + c)}
\]

\[
\frac{\partial \Delta \text{FMQ}}{\partial \text{AMCNK}} = -\left[\log_{10}(D) - b \cdot \text{AMCNK} + d \cdot (a \cdot \text{AMCNK} + c)\right] \cdot \frac{1}{a \cdot \text{AMCNK} + c}
\]

\[
\frac{\partial \Delta \text{FMQ}}{\partial \text{Al}_2\text{O}_3} = \frac{1}{\text{Denom}} \cdot \frac{\partial \Delta \text{FMQ}}{\partial \text{AMCNK}}
\]

\[
\frac{\partial \Delta \text{FMQ}}{\partial \text{Denom}} = -\frac{\text{Al}_2\text{O}_3}{\text{Denom}} \cdot \frac{\partial \Delta \text{FMQ}}{\partial \text{AMCNK}}
\]

b) For the uncertainty inherent to the model:

\[
\frac{\partial \Delta \text{FMQ}}{\partial a} = \frac{\log_{10}(D) - b \cdot \text{AMCNK} + d}{(a \cdot \text{AMCNK} + c)^2} \cdot \text{AMCNK}
\]

\[
\frac{\partial \Delta \text{FMQ}}{\partial b} = -\frac{a \cdot \text{AMCNK}}{(a \cdot \text{AMCNK} + c)^2}
\]

\[
\frac{\partial \Delta \text{FMQ}}{\partial c} = -\frac{1}{a \cdot \text{AMCNK} + c}
\]

From the individual partial derivatives and variances ($\sigma^2$) of FeO$_{\text{tot mgt}}$, FeO$_{\text{tot melt}}$, TiO$_2$$_{\text{mgt}}$, TiO$_2$$_{\text{melt}}$, Al$_2$O$_3$, Denom, a, b, c and d the overall uncertainty can be estimated according to Equation S-2. For the estimation of the propagated analytical uncertainty only the partial derivatives and variances of FeO$_{\text{tot mgt}}$, FeO$_{\text{tot melt}}$, TiO$_2$$_{\text{mgt}}$, TiO$_2$$_{\text{melt}}$, Al$_2$O$_3$ and Denom have to be considered.

**Figure S-1** Dependence of the Fe–Ti exchange coefficient between magnetite and rhyolitic melt on $f$O$_2$ and melt alumina saturation index. $D_{\text{Fe-Ti}}$$_{\text{magnetite}}$ refers to $(D_{\text{FeO}_{\text{tot}}}_{\text{magnetite}})/(D_{\text{TiO}_2}_{\text{magnetite}})$, whereas ASI refers to molar Al$_2$O$_3$ / (CaO + Na$_2$O + K$_2$O). The dataset comprises magnetite–melt pairs from 50 different experiments performed at three different oxygen fugacity buffers, temperatures of 800-1000 °C, pressures of 100-500 MPa, with melt ASI values of 0.71-1.12, and magnetite compositions of 0.2-14 wt. % TiO$_2$ (Arató and Audétat, 2017).
**Figure S-2** Variance of the Mn–Mg exchange coefficient between magnetite and silicate melt as a function of the MnO/MgO weight ratio in the silicate melt. $D_{\text{Mn–Mg}}^{\text{mgt/melt}}$ refers to $(D_{\text{MnO}}^{\text{mgt/melt}})/(D_{\text{MgO}}^{\text{mgt/melt}})$, with MnO and MgO given in weight percent. The data set comprises the 296 magnetite–rhyolite pairs from own experiments, plus the data points shown in Figure 2. The shaded envelope, which encompasses $D_{\text{Mn–Mg}}^{\text{mgt/melt}}$ values of 1-4, comprises 95 % of all data points.

**Figure S-3** Comparison of reported experimental $fO_2$ values with $fO_2$ values obtained via magnetite–ilmenite (Ghiorso and Evans, 2008) for 59 ilmenite-bearing experiments taken from the literature. Oxygen fugacities are reported in log units relative to the fayalite-magnetite-quartz (FMQ) buffer.
Figure S-4 Difference between the $f_O^2$ values calculated via magnetite-ilmenite and ones calculated via FeTiMM [$\Delta f_O^2 (\text{FeTiMM} - \text{mgt--ilm})$], as a function of (a) $f_O^2$, (b) AMCNK, (c) temperature, (d) magnetite composition, and (e) melt SiO$_2$ content, for the 109 data points that were used to calibrate FeTiMM.
Figure S-5 Difference between reported experimental $f_{O_2}$ and $f_{O_2}$ calculated via FeTiMM [Δ$f_{O_2}$ (FeTiMM - experimental)], as a function of (a) $f_{O_2}$, (b) AMCNK, (c) temperature, (d) magnetite composition, and (e) melt SiO$_2$ content, for 27 ilmenite-undersaturated experiments taken from the literature.
Supplementary Tables

Supplementary Tables S-1 and S-2 can be downloaded as Excel files at http://www.geochemicalperspectivesletters.org/article1740.

Table S-1  Full data set used to calibrate and test the FeTiMM model. The references listed in this table are detailed in Table S-2.

Table S-2  List of experimental studies that were screened for the five criteria discussed in the text.

Supplementary Information References

