The isotope composition of natural material can vary, either through the decay of a radioactive parent that results in radiogenic ingrowth of a particular isotope, or as a product of chemical reactions driven by physical changes, for example, during the interactions between biosphere, hydrosphere, and rocks (1). Stable-isotope fractionation can reach levels of several percent for light elements, such as hydrogen, carbon, nitrogen, oxygen, or sulfur. However, the magnitude of isotope fractionation drastically decreases with the nuclear mass \( M \) (as \( \sim 1/M^2 \)) and also with temperature. Accordingly, magmatic fractionation of (heavy) metal isotopes was long considered to be insignificant. Early studies on samples that formed at high temperatures, such as meteorites, used the isotopic composition of metals [such as iron (Fe), copper, and others] to detect heterogeneities from the origin of our solar system or to address processes of planetary accretion [see, for example, (2–4)]. However, the study by Teng et al. on page 1620 in this issue (5) strongly indicates that Fe-isotope fractionation during magmatic differentiation needs to be considered when investigating planetary materials.

A remarkable range of Fe-isotope variations [on the order of \( \sim 1 \) per mil (‰)] in high-temperature environments, such as Earth’s mantle, was first observed by Williams et al. (6, 7). Their findings could not be explained by the recycling of material into the mantle that had isotopically fractionated previously during low-temperature processes on Earth’s surface. Rather, their results indicated that these isotopic variations in the mantle were produced by metasomatic processes, in which alteration is driven by interactions of the mantle with small amounts of melts or fluids.

Subsequently, Weyer et al. (8, 9) observed that Fe appears to be isotopically lighter in Earth’s mantle than in the crust, by \( \sim 0.1\% \). Additionally, Weyer and Ionov (10) observed that in several suites of mantle rocks, Fe-isotope fractionation was coupled with the amount of melt that was extracted from these rocks.

These findings indicated that Fe isotopes can fractionate during partial melting in the mantle, at temperatures of 1200°C or higher (see the figure). However, because most mantle rocks that were brought to Earth’s surface originate from the uppermost mantle, which is commonly modified by fluids and recycled material, it was still highly debated whether these isotopic differences between mantle and crust (and isotopic trends in the mantle) stem from partial melting or from metasomatic processes (8, 11, 12). This question is intriguing, because isotope fractionation during magmatic processes has not yet been observed for any other metal.

Teng et al. now provide convincing evidence that Fe isotopes fractionate during magmatic differentiation. Because mantle rocks in equilibrium with melt are difficult to find, these authors studied the opposite process, the fractional crystallization of olivine from a magma. They investigated the Fe-isotope composition of a suite of basalts from a lava lake in Hawaii and that of corresponding olivine grains that crystallized from the lava.

They observed that Fe in basalts becomes isotopically heavier as more olivine has crystallized, and that olivines are always isotopically lighter than the coexisting basaltic melts from which they formed (see the figure).

Studies similar to the investigations of Teng et al. have been performed already for lighter metal isotope systems, such as lithium (Li) and magnesium (Mg) (13, 14). However, no isotopic differences between basalts and olivine crystals were observed. Why do the isotopes of Fe fractionate during magmatic...
Fates of iron isotopes. Schematic of Fe-isotope fractionation during magmatic processes. During partial melting in the mantle, the heavy Fe isotopes preferentially enter the melt, resulting in high $^{56}\text{Fe}/^{54}\text{Fe}$ in melts and the crust, and low $^{56}\text{Fe}/^{54}\text{Fe}$ in the depleted mantle (the latter does not differ much from that of the bulk silicate Earth, for mass-balance reasons). Likewise, light Fe isotopes preferentially enter olivine during fractional crystallization, resulting in low $^{56}\text{Fe}/^{54}\text{Fe}$ in the crystals and increasingly higher $^{56}\text{Fe}/^{54}\text{Fe}$ in the remaining melt. Fractional crystallization of magnetite has the opposite effect, however (15).

processes, when other metal ions do not? There is no definitive answer to this question yet. However, in contrast to Li and Mg, Fe occurs in two different oxidation states in basalts ($\text{Fe}^{2+}$ and $\text{Fe}^{3+}$). Only $\text{Fe}^{2+}$ fits into the olivine structure, whereas $\text{Fe}^{3+}$ preferentially stays in the melt. Potentially, this different partitioning of Fe species leads to measurable fractionation of their isotopes. Future experimental studies may provide the answer to this question, and also resolve whether this isotope fractionation occurs as an equilibrium or kinetic process.

In addition, because olivine is the most abundant mineral in the upper mantle and is the major host for Fe, we can speculate that Fe-isotope fractionation during melting in the mantle must occur as well. Indeed, the direction of isotope fractionation between olivine and melt during fractional crystallization agrees with that earlier predicted to occur during partial melting ($8–10$); that is, olivine and mantle rocks are isotopically lighter than basalt.

These findings of magmatic Fe-isotope fractionation open new possibilities for studying magmatic processes on Earth and other planets. Future studies may link the small differences we are observing in the Fe-isotope composition of materials from different planets—for example, the apparently heavier Fe-isotope compositions of lunar and terrestrial basalts compared to the planets Mars and Vesta ($3, 8, 9$)—to their respective conditions during planetary differentiation.

References