make the matrix and is repressed by SinR.

How does EpsE work? A search for mutants that could swim even with EpsE expressed found several mutations in *fliG*, a gene distant from the matrix-encoding operon that codes for a protein involved in torque generation in the flagellar rotor. Induced expression of EpsE in *B. subtilis* stopped cell motility. Fluorescent-labeled EpsE localized at spots corresponding to individual motors, suggesting direct interaction with FliG, but not in cells with the mutations rendering FliG insensitive to EpsE.

To determine whether EpsE acts as a brake that locks the motor, or a clutch that leaves the rotor freely spinning, Blair *et al.* tethered bacteria to a substrate by their filaments and observed rotation of the cell bodies around single flagellar motors. Under the influence of EpsE, cells stopped spinning but continued to undergo free rotational Brownian motion, indicating a clutch mechanism.

The direct inhibition of motor rotation by EpsE represents a newly discovered control mechanism for bacterial swimming. Bacterial flagella are large protein complexes that require about 40 to 50 genes to assemble (*6*). Thus, the most obvious advantage of the EpsE mechanism over transcriptional control of flagellar genes is speed. In *B. subtilis*, only one protein, EpsE, needs to be expressed to stop the motor. Presumably, this is important if cells are to stay put in the early stages of biofilm formation. However, the advantages of a clutch over a brake mechanism are not so clear. Perhaps free rotation of flagella—or, alternatively, reduced motility during the transition to the EpsE-inhibited state—is important for the formation of well-structured biofilms (*7*). Or maybe a clutch is simply easier to make than a brake.

Whether this is a universal mechanism or a peculiarity of *B. subtilis* remains to be discovered, as do the details of how the clutch works and how it helps biofilm formation. More experiments like this are needed, not just to find genes and proteins, but to learn what they do and how.

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GEOCHEMISTRY

What Drives Iron Isotope Fractionation in Magma?

Unlike other metals, magmatic melting and recrystallization fractionate iron isotopes, possibly because of the different oxidation states of iron.

Stefan Weyer

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 ~$1/M^2$) 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 ≈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 ~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

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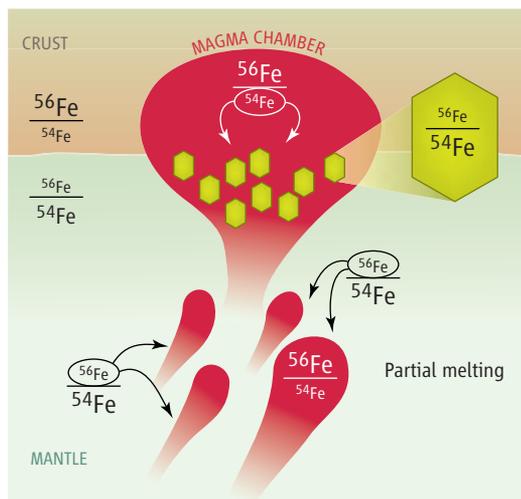
Fates of iron isotopes. Schematic of Fe-isotope fractionation during magmatic processes. During partial melting in the mantle, the heavy Fe isotopes preferentially enter the melt, resulting in high $^{56}Fe/^{54}Fe$ in melts and the crust, and low $^{56}Fe/^{54}Fe$ in the depleted mantle (the latter 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 $^{56}Fe/^{54}Fe$ in the crystals and increasingly higher $^{56}Fe/^{54}Fe$ in the remaining melt. Fractional crystallization of magnetite has the opposite effect, however (*15*).

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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 those from the planets Mars and Vesta (*3, 8, 9*)—to their respective conditions during planetary differentiation.

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 (Fe^{2+} and Fe^{3+}). Only Fe^{2+} fits into the olivine structure, whereas 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

APPLIED PHYSICS

Diamond for Quantum Computing

Optically active defects in diamonds are promising candidates for the building blocks of quantum computers.

Steven Prawer and Andrew D. Greentree

For the technologist seeking to build devices that take advantage of the quantum mechanical properties of coherence and entanglement, diamond looks to be the ideal material. Single-crystal diamond has long held allure as a gemstone, and its extreme electrical, optical, and mechanical properties have already found applications such as heat spreaders, optical windows, electrodes for electrochemistry, high-energy particle detectors, dosimeters, and biosensors. But it is in the quantum realm that diamond truly stands apart, its optical properties tailor-made for the fabrication of the building blocks of new quantum technologies. The optical centers in diamond offer access to iso-

lated quantum systems that can be controlled at room temperature.

A color center is an impurity or defect in a crystal and is responsible for the colors of emeralds and rubies. The color center most used in diamond is the negatively charged nitrogen vacancy (NV) center. These optically active centers, consisting of a substitutional nitrogen atom next to a missing carbon atom (see the figure, left panel), are so bright that they can be detected individually with conventional microscopy. Because the ground state shines more brightly than the excited state, the state of the NV center can be read out. Resonant microwave pulses allow full quantum control of the state of the center.

These properties have enabled the construction of the most basic building block of a quantum processor, the quantum bit or qubit that is operable at room temperature. This is

revolutionary in terms of solid-state quantum computing, where the usual temperature scales being discussed are fractions of a kelvin. The time required to manipulate the state of the qubit is brief (tens of nanoseconds), and the measured room-temperature decoherence rate is measured to be 0.35 ms (*1*), meaning that some 10^4 operations can be performed before decoherence takes over and the state is lost. Demonstrations of multiqubit couplings (albeit in a nonscalable design) (*1–3*) show promise for building small quantum memories and other few-qubit protocols.

When a single atom deexcites, it emits a single photon, and these photons can be used for a number of applications including quantum metrology, imaging, and ultrasecure communications using quantum key distribution (QKD). The NV centers in diamond mimic these single atoms, displaying photostable

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