important for regulation of PEPCK expression. Any broad conclusions, however, must await further studies on other key gluconeogenic enzyme genes and their expression in response to perturbations other than birth, such as fasting and refeeding.

Why is it that mice with defective eIF2α phosphorylation exhibit both β cell insufficiency and defective liver gluconeogenesis, whereas PERK-deficient animals only exhibit β cell insufficiency? The Harding *et al.* work (*1*) provides a possible answer: PERK-deficient mice have a reduced ratio of phosphorylated to total eIF2α in pancreas, lung, and thymus, but a normal ratio in liver and spleen. This finding suggests that an eIF2α kinase other than PERK may be the key modulator of translational control of gluconeogenic enzyme expression in the liver. If eIF2α phosphorylation does prove to be important in this pathway, the mammalian homolog of yeast *GCN2* (which is activated by amino acid deprivation) may be involved (see the figure). Defective gene expression downstream of GCN2, however, is unlikely to account for all of the characteristics of the eIF2 mutant mice, given that *GCN2*-deficient animals do not manifest any impairment in neonatal survival (*12*).

The importance of the new mouse models is underscored by the discovery of mutations in the *PERK* gene in an inherited autosomal recessive disease in humans, called the Wolcott-Rallison syndrome (*13*). This disease is classified as a form of type 1 diabetes because it develops in early infancy and is characterized by the destruction of pancreatic β cells. A key question posed by the Harding and Scheuner studies is why β cells are selectively destroyed in Wolcott-Rallison patients and in the mutant mice. It is very likely that β cells die because they need both PERK (which is extraordinarily abundant in the pancreas) and eIF2α phosphorylation to survive.

In yeast, eIF2α is phosphorylated in response to nutritional cues and directs adaptations in intermediary metabolism. The new work suggests that, despite considerable diversification in upstream signals, metazoans have retained the kinases that phosphorylate eIF2α and control translation, adapting them for the regulation of glucose homeostasis.

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PERSPECTIVES: EARTH HISTORY

The Rise of Atmospheric Oxygen

James F. Kasting

Geologists and evolutionary biologists have long speculated about when atmospheric oxygen, O_2, first became abundant and how rapidly it reached its present concentration. Cloud's groundbreaking studies of reduced and oxidized iron and uranium minerals (*1*), coupled with later studies of microfossils and ancient soils (*2*), convinced most observers that atmospheric O_2 underwent a dramatic increase in concentration between 2200 and 2400 million years ago. But some geologists remained skeptical (*3*) because the geochemical and biological data allowed room for alternative interpretations.

Recently, an entirely new line of geochemical evidence removed much of the uncertainty regarding the timing of the initial rise in atmospheric O_2. Farquhar *et al.* (*4*) showed that sulfur isotope ratios in rocks older than about 2300 million years are unusual. In younger rocks, the difference in abundance between ^{33}S and ^{32}S is about half that between ^{34}S and ^{32}S. Such "mass-dependent" isotope fractionation results from a variety of aqueous chemical and biochemical reactions. But the sulfur isotopes in rocks older than 2300 million years consistently deviate from the mass-dependent fractionation line. The "mass-independent" fractionation seen in these rocks is thought to result solely from gas-phase photochemical reactions such as photolysis of SO_2 (*4*).

In a high-O_2 atmosphere like the present one, the isotopic signature of such reactions is unlikely to be seen in sediments because almost all sulfur gases emitted into the atmosphere are oxidized to sulfuric acid and accumulate in the ocean as dissolved sulfate, SO_4^-. In a low-O_2 atmosphere, however, sulfur can exit the atmosphere in a variety of oxidation states from −2 (H_2S) to +6 (H_2SO_4) (see the figure) (*5*). The likelihood of transferring a mass-independent fractionation pattern to sediments is thus much higher in a low-O_2 atmosphere. Two independent sets of sedimentary S isotope data (*6*, *7*) agree with those of Farquhar *et al.* (*4*), strongly supporting the hypothesis that the transition from an anoxic to an oxic atmosphere occurred near 2300 million years ago.

The timing of the initial O_2 rise is thus relatively well established, but the question of what triggered it remains hotly debated. Researchers agree that O_2 was produced initially by cyanobacteria, the only prokaryotic organisms (Bacteria and Archea) capable of oxygenic photosynthesis. But cyanobacteria are thought to have emerged by 2700 million years ago, on the basis of evidence from organic biomarkers in well-preserved sedimentary rocks (*8*). [Earlier claims for the existence of cyanobacteria 3500 million years ago (*9*) have recently been called into question (*10*).] Why was there a gap of at least 400 million years between the emergence of cyanobacteria and the rise in atmospheric O_2?

A clue may come from the carbon isotope composition of sedimentary carbonates, which has remained essentially constant through most of the geologic record. This suggests that organic carbon was buried at a more or less constant rate since at least 3200 million years ago (*11*). If this organic carbon was produced by photosynthesis, then net O_2 production would have been constant as well (or, more precisely, it would have remained proportional to the rate of volcanic CO_2 emission). If so, the sink for oxygen must have changed rather than its source. This in turn suggests a geological rather than a biological explanation for the timing of the rise in O_2.

On page 839 of this issue, Catling *et al.* (*12*) suggest that the transition from low to high O_2 was caused by enhanced hydrogen escape into space as a result of high methane concentrations in the Late Archean/Early Proterozoic atmosphere (3000 to 2300 million years ago). As was realized a long time ago (*13*), hydrogen should escape rapidly from low-O_2 atmospheres rich in reduced gases such as H_2 and CH_4. The vast majority of Earth's hydrogen was originally incorporated into the planet as water. The escape of hydrogen to space thus results in a net accumulation of oxygen somewhere on the planet. Kasting *et al.* have argued (*14*) that most of the escaping hydrogen originated from

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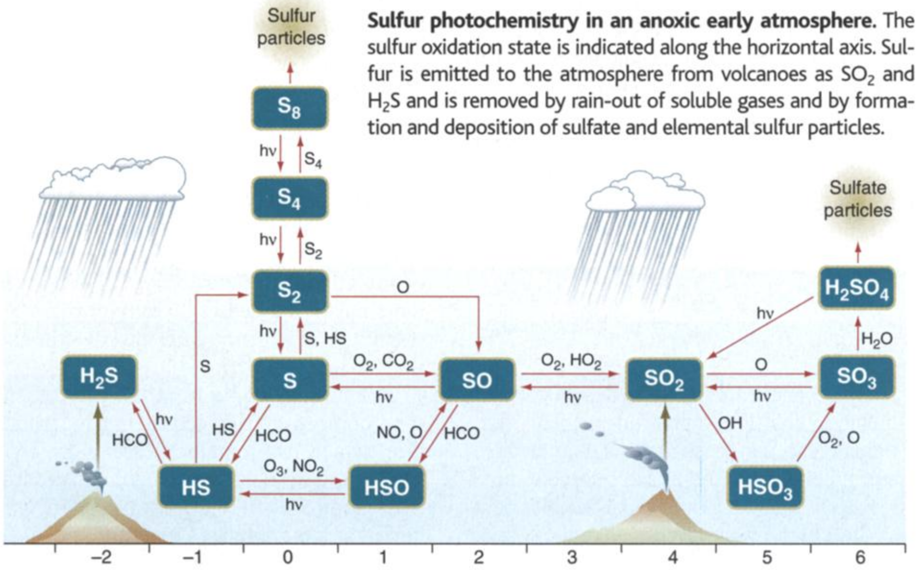
volcanic outgassing of H_2 and that the oxygen left behind caused a gradual increase in the oxidation state of the upper mantle. Volcanic gases became more oxidized with time, culminating in a transition to an O_2-rich atmosphere once net photosynthetic production of O_2 exceeded the volcanic sink. This idea appears to have been incorrect. Data on the V and Cr content of ancient basalts and komatiites (special high-MgO igneous rocks) indicate that the oxidation state of the mantle has not changed appreciably with time (*15*).

Catling *et al.* have added a couple of new twists to this argument that may help to revive it, albeit in a somewhat different form. First, they suggest that the hydrogen escape rate may have been greatly enhanced by the buildup of biogenic methane in the Archean atmosphere. They suggest that most of this methane was produced by anaerobic decomposition of photosynthetically produced organic matter. Thus, the overall reaction was $CO_2 + 2H_2O \rightarrow CH_4 + 2O_2$. If O_2 was consumed faster than CH_4—which must have been true in order for the atmosphere to remain anoxic—then atmospheric CH_4 could have reached concentrations of 0.1% or more. This scenario is consistent with evidence that the Late Archean climate was warm despite low atmospheric CO_2 concentrations (*16*). CH_4 is a good greenhouse gas and could have supplied the necessary warming (*17*).



Sulfur photochemistry in an anoxic early atmosphere. The sulfur oxidation state is indicated along the horizontal axis. Sulfur is emitted to the atmosphere from volcanoes as SO_2 and H_2S and is removed by rain-out of soluble gases and by formation and deposition of sulfate and elemental sulfur particles.

Second, Catling *et al.* propose that the oxygen left behind by escape of hydrogen to space was largely consumed by oxidation of continental crust. This, in turn, caused a change in the oxidation state of gases released by low-temperature metamorphic processes. Again, the transition to an oxic atmosphere occurs when the net photosynthetic production of O_2 exceeds the input of reduced gases to the atmosphere. However, the oxidation state of the mantle remains unchanged, thus removing the conflict with the V and Cr data.

Catling *et al.* have explored their hypothesis from a number of different angles and have presented a good case. However, some details of their model remain to be filled in. Perhaps most importantly, how did the oxygen produced by photosynthesis get incorporated into the continents? This could not have occurred by direct oxidative weathering because that is contradicted by the persistence of detrital uraninite and pyrite in Archean sediments (*1*). Some of the dissolved iron that was oxidized in the ocean and deposited in banded iron formations (BIFs) was added to the continents, but most of it probably ended up in the mantle (*12*). The release of reduced metamorphic gases would have oxidized the continental crust only if those gases were generated from oxidized species such as H_2O and CO_2 and not from organic matter. This requires low-temperature crustal oxidation reactions that are not described by the authors.

Indeed, the changing nature of crustal oxidation processes may be the key to the whole problem. Keeping atmospheric O_2 concentrations low before 2300 million years ago in the face of modern rates of O_2 generation requires mechanisms for oxidizing crustal materials that did not involve free O_2 or sulfate, both of which are thought to have been scarce.

One such process that occurs on the sea floor today is "serpentinization." Magnesium-rich basalts form serpentine minerals during low-temperature aqueous alteration (*18*), often in fracture zones along slow-spreading mid-ocean ridges (*19*). In this process, ferrous iron released from the basalt is partially oxidized to form magnetite. When water is the oxidant, the overall reaction is $3FeO + H_2O \rightarrow Fe_3O_4 + H_2$. In the presence of CO_2, CH_4 is produced instead. Most of these rocks are subducted back into the mantle, but that may not matter from the standpoint of atmospheric oxidation balance. Serpentinization generates reduced gases. If the Archean sea floor was more mafic (that is, rich in iron and magnesium and poor in silica) as a result of higher mantle temperatures and more partial melting beneath mid-ocean ridges, the generation rates of H_2 and CH_4 might have been substantially higher than today. Alternatively, higher mantle temperatures could have led to shallower mid-ocean ridges and hotter, two-phase hydrothermal vent fluids, both of which would have promoted the release of dissolved ferrous iron (*20*). Oxidation of this iron to form banded iron formations might then have been a bigger O_2 sink than has been previously estimated (*21*).

Perhaps these two processes together sequestered enough oxygen to suppress atmospheric O_2 before 2300 million years ago. The rise of O_2 at this time might then be linked to the gradual cooling of Earth's interior. As suggested by Catling *et al.* (*12*), a sudden rise in atmospheric O_2 would have occurred when the flux of reduced gases fell below the net photosynthetic production rate of oxygen. A search for trends in the composition and oxidation state of ancient oceanic and continental crust might allow us to determine which of these various processes was most important.

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