The Global Phosphorus Cycle: Past, Present, and Future

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The cycling of phosphorus, a biocritical element in short supply in nature, is an important Earth system process. Variations in the phosphorus cycle have occurred in the past. For example, the rapid uplift of the Himalayan-Tibet Plateau increased chemical weathering, which led to enhanced input of phosphorus to the oceans. This drove the late Miocene "biogenic bloom." Additionally, phosphorus is redistributed on glacial timescales, resulting from the loss of the substantial continental margin sink for reactive P during glacial sea-level lowstands. The modern terrestrial phosphorus cycle is dominated by agriculture and human activity. The natural riverine load of phosphorus has doubled due to increased use of fertilizers, deforestation and soil loss, and sewage sources. This has led to eutrophication of lakes and coastal areas, and will continue to have an impact for several thousand years based on forward modeling of human activities.

KEYWORDS: phosphorus, biogeochemistry, soil, cycling, paleoclimatology

IMPORTANCE OF PHOSPHORUS—WHY P?

Nature apparently made an odd choice when it chose to build most of the critical photosynthesis and metabolic pathways around phosphorus (Westheimer 1987). Phosphorus (P) is present in only minute quantities in Earth's crust (0.09 wt%). It is interesting to note that P is twice as abundant in Martian rocks (Greenwood and Blake 2006). Once P is liberated from minerals during weathering, it is quickly sequestered into a number of more recalcitrant phases, limiting its day-to-day accessibility to plants and organisms. Phosphorus has no stable atmospheric gas phases, so unlike the case for other critical bionutrients, such as nitrogen and carbon, ecosystems have to depend on aqueous transfer of this nutrient. The only exceptions are completely P-depleted ecosystems. Examples include transport-limited regimes such as the Amazon Basin, and old volcanic islands like Kauai (e.g. Kurtz et al. 2001) where ecosystem survival depends on the minute amount of P that is delivered by trans-Pacific dust. In summary, there simply is not much P around, it does not travel well, and it is not readily bioavailable.

In spite of these limitations, P plays an essential role in biological systems. We and all other vertebrates owe our gravity-defying skeletons to the hydroxylapatite that makes up bones (for more on P mineral chemistry, see Oelkers and Valsami-Jones 2008 this issue). We also depend on hydroxylapatite (with carbonate and fluoride substitution) to make our teeth strong enough to bite into a crispy apple. Growing that apple, in turn, depended on the Calvin cycle photosynthetic pathway and the important role that adenosine triphosphate (ATP) plays in this cycle. The double helix of our DNA is only possible because of the phosphate ester bridges that bind the helix strands. All cells owe their very structure to the phospholipids that make up cell wall membranes. Earth's biological systems have depended on P since the beginning of life (e.g. Nealson and Rye 2004), and we have no reason yet to doubt that P is critical to all biological systems in the universe. The unique energetics of the phosphate molecule are central to the function of ATP, the linchpin for metabolism in biological systems. ATP is the most abundant biomolecule in nature (Schlesinger 1997).

Phosphate hydrolyzes only slowly in the absence of enzymes. It is very stable at the pH of cell interiors, but it hydrolyzes rapidly when enzymes are present. Phosphate mobility within organisms is high, and thus in contrast with environmental conditions, phosphate is readily available within biological systems. Finally, P is the essential ingredient in bone, contributing to a hydroxylapatite/organic scaffold that has strength, some flexibility, an open structure for substitution of some essential nutrients, and a biologically friendly dissolution and precipitation behavior, i.e. it rapidly dissolves and precipitates when solution composition is right (Skinner and Jahren 2004). Thus, in spite of low natural abundance, the chemical behavior of phosphate makes it the most likely candidate around which to build biological systems.

In this paper, I explore the sources, interactions, and eventual fate of P on Earth from the system perspective. I also outline efforts to understand variations in P cycling, with a look both forward and backward through time. As we have gained more and more understanding of various aspects of global P cycling, we have put some debates to rest and have sparked others. But one aspect has remained constant: an understanding of the P cycle is critical to our understanding of biological, chemical, and geological cycles on Earth and beyond, so it will likely continue to be an active research topic for decades to come.

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THE GLOBAL PHOSPHORUS CYCLE IN THE CONTEXT OF THE PREHUMAN ERA

Phosphorus is present at an appreciable concentration in only a few minerals. The main P-bearing mineral in igneous rocks is fluorapatite, an early-formed accessory mineral. It appears as tiny, euhedral crystals associated with ferromagnesian minerals. In sedimentary rocks, P is typically associated with authigenic carbonate-fluorapatite (CFA; more on this later). All apatite minerals contain phosphate oxyanions linked by Ca^{2+} cations to form a hexagonal framework, but they contain different elements at the corners of the hexagonal cell.

Apatite minerals weather congruently (that is, the mineral weathers completely to dissolved products in one step) as a result of reaction with dissolved carbon dioxide in the form of carbonic acid:

$$Ca_5(PO_4)_3OH + 4H_2CO_3 \leftrightarrow 5Ca^{2+} + 3HPO_4^{2-} + 4HCO_3^- + H_2O$$
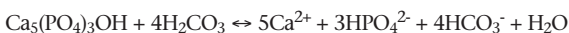
In soils, weathering releases P from minerals by several processes. First, biochemical respiration releases CO_2, increasing acidity around degrading organic matter and root hairs; poorly crystalline P-bearing minerals dissolve rapidly in this acidic environment, releasing P to root pore spaces (e.g. Schlesinger 1997). Second, organic acid exudates from plant roots also can dissolve apatite and release P to soil pore spaces (Schlesinger 1997). Upon dissolution, P is transformed by several processes into forms that are less immediately bioavailable. It can be incorporated into plant tissue where it is converted into an organic form. Plants actively incorporate phosphate to form biomolecules critical to photosynthesis and metabolism. Plants have developed several tactics to increase the supply of P to roots, including the secretion of phosphatase, an enzyme that can release bioavailable inorganic P from organic matter into soil pore spaces (Bucher 2007). The symbiotic fungi *mycorrhizae* can coat plant rootlets, excreting phosphatase and organic acids to release P and providing an active uptake site for the rapid diffusion of P from soil pore spaces to the root surface. In exchange, the plant provides carbohydrates to the mycorrhizal fungi (Smith et al. 2003). When plants or tissues die, organically bound phosphorus undergoes the same fate as organic litter; bacterial and fungal reactions in soils progress slowly, oxidizing organic matter and releasing P as phosphate to soil solutions (Bucher 2007).

An additional sink for P released by weathering is iron and manganese oxyhydroxides, where P is either co-precipitated with them or adsorbed onto their surfaces during soil weathering. These reducible Fe and Mn species are termed "occluded" fractions in soils. They have a large binding capacity for phosphate because of their extremely high surface area and their overall positive charge (Filippelli 2002). These P-bearing phases often constitute the main, long-term storage pool for soil P. The labile (unstable) forms include P in soil pore spaces (as dissolved phosphate ion) and adsorbed onto soil particle surfaces (these forms are termed "non-occluded"), as well as P incorporated in soil organic matter. On a newly exposed rock surface, nearly all of the P is in apatite. With time and soil development, however, it is released and incorporated in other phases (Fig. 1; Filippelli et al. 2006). Over time, the total amount of P available in the soil profile decreases, as soil P is lost through surface and subsurface runoff (e.g. McDowell and Sharpley 2001). Eventually, the system reaches a terminal steady state, in which soil P is heavily recycled and any P lost through runoff is balanced by new P weathered from apatite at the base of the soil column.
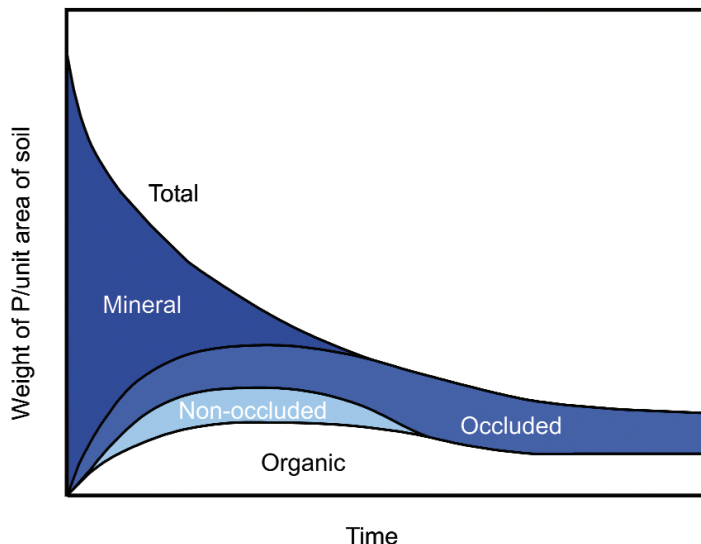

FIGURE 1 Modeled changes in soil phosphorus availability with time (after Filippelli 2002), showing transformation of mineral phosphorus into non-occluded (i.e. highly labile) and organic forms before eventual dominance of occluded (oxyhydroxide-bound) and organic forms. The bioreactivity of phosphorus increases from mineral to occluded to organic forms. Note that phosphorus is continuously lost from the system.

Transport from Land to Sea

Transport of eroded soil by rivers delivers P to the oceans. Riverine P is in two main forms: particulate and dissolved. Most of the particulate P in rivers is held within mineral lattices and never participates in the active biogenic cycle. This is also true after delivery to the oceans, because dissolution rates in seawater, where pH is high and ionic buffering is strong, are exceedingly low. Thus, much of the P physically eroded from continents is delivered relatively unaltered to the oceans, where it is sedimented on continental margins and in the deep sea, waiting for subduction or accretion to give it another chance to participate in the continental P cycle. Some P-bearing particulate is adsorbed onto the surfaces of other soil particles, some P is held within soil oxyhydroxides, and some is incorporated into particulate organic matter. The fate of organic P after transfer to the ocean is poorly understood. For example, P adsorbed onto soil surfaces may be effectively removed in response to the high ionic strength of ocean water, providing an additional phosphate source to the ocean. A small amount may be released from terrestrial organic matter during bacterial oxidation after sediment burial. Finally, some sedimentary environments along continental margins are suboxic or even anoxic, conditions that favor oxyhydroxide dissolution and release of sorbed P. Several important studies have examined the transfer of P between terrestrial and marine environments (e.g. Ruttenberg and Goñi 1997), but quantification of the interactions between dissolved and particulate forms and the aquatic/marine interface is lacking.

Marine Cycling

Once in the marine system, dissolved P is a limiting nutrient for biological productivity (e.g. Ammerman et al. 2003) and is perhaps *the* ultimate limiter of ocean productivity on geologic timescales (Tyrrell 1999; Bjerrum and Canfield 2002). Phosphorus concentrations follow the nutrient profile in the ocean, with surface depletion and deep enrichment. Concentrations are near zero in most surface waters, as P is taken up by phytoplankton as a vital component of

their photosystems. In deep water, phosphate concentration increases with water age, so values in the young, deep waters of the Atlantic are 40% lower than those in the older Pacific Ocean. Once P is incorporated into organic matter, it follows a similar biogeochemical route as the organic matter itself, undergoing active recycling in the water column and at the sediment–water interface. Several recent examinations of P cycling and recycling (e.g. Benitez-Nelson 2000; Paytan and McLaughlin 2007) elucidate the process in which organic P in the water column is transferred from organisms to dissolved inorganic forms and back to organisms.

Sedimentary Burial and Diagenesis

Recent research on the phosphorus geochemistry of deep-sea and continental-margin sediments has revealed several commonalities in burial dynamics. First, the most important shuttle of reactive P to the seafloor in most settings is organic matter. Organic P may be released during oxidation in the water column, at the sediment–water interface, and after burial. Some studies point to enhanced P release, at the expense of organic C, during organic matter degradation under suboxic and anoxic conditions (e.g. Colman and Holland 2000; Schenau et al. 2000; Filippelli 2001), providing a positive feedback mechanism for enhanced productivity should the released P reach the photic zone (layer of the ocean that is penetrated by sunlight and where photosynthesis can occur). This has been used to explain sedimentary sequences associated with sapropels (sediments rich in organic matter) and oceanic anoxic events (Van Cappellen and Ingall 1996; Filippelli et al. 2003; Handoh and Lenton 2003; Mort et al. 2007). If P release from organic matter occurs in the water column or near the sediment–water interface, this released P may return to the reactive marine pool. Increasing evidence from a variety of marine sedimentary settings, however, indicates that as P is released below the sediment–water interface, it can reach concentrations in the pore water sufficient to induce the precipitation of CFA. This process explains the mirror-image trends of decreasing organic P and increasing authigenic P with depth that are observed in a wide range of settings (FIG. 2). The net result of mineral trapping of reactive P is that it becomes permanently stored in marine sediments, resistant to further release until the sediment is either subducted, with P eventually returning to the Earth's surface in an igneous mineral, or uplifted by tectonic processes, releasing the CFA-bound P during subaerial weathering (Guidry and Mackenzie 2003).

THE GLOBAL PHOSPHORUS CYCLE IN THE PAST: A RECORD OF GLOBAL CHANGE

Perhaps the most important questions to ask about the marine P cycle include how it has varied in the past, what has driven this variation, and how this has affected other biogeochemical systems. Although simple to ask, these questions are very difficult to answer when viewed through the filter of the geologic record. For example, the most recent paradigm for the episodic nature of large marine deposits rich in P, called phosphorites and typified by the massive Permian Phosphoria Formation of the western US, was that these deposits reflect times in Earth's history when the transport of P from continental weathering to the ocean was significantly higher than it is today. What had not been recognized in this argument was that phosphorite deposits are the result of intervals of "normal" marine organic matter deposition in regions of high biological productivity on continental margins and that P accumulation rates were similar to those of today. Deposition was followed by intervals of significant sediment reworking (often in response to sea level variations) and concentration of the disseminated P
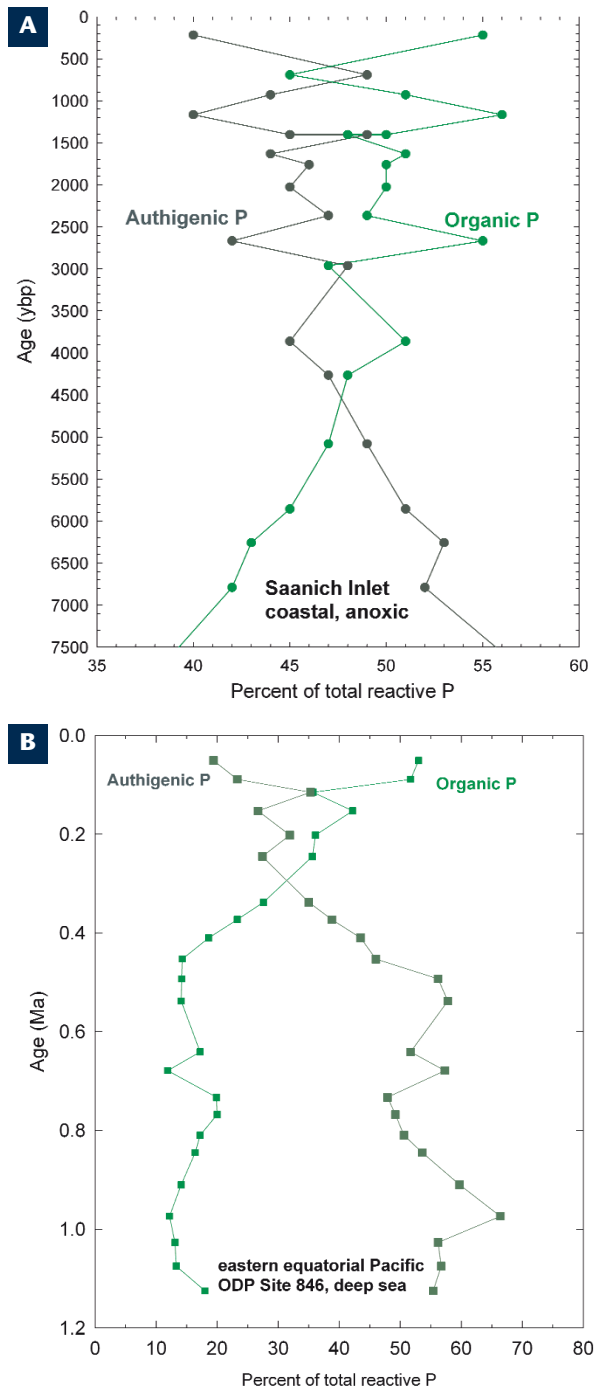
FIGURE 2 Percent of total reactive P in authigenic and organic phases for (**A**) the anoxic Saanich Inlet (ODP Site 1033) and (**B**) the deep-sea eastern equatorial Pacific (ODP Site 846). This comparison reveals the sink-switching that occurs during sedimentary P cycling, with the growth of an authigenic P phase and loss of P from organic phases with age and depth.

into phosphatic lags and hard grounds. An example of a high-P sedimentary rock from the Miocene Monterey Formation is presented in FIGURE 3. When this process repeats itself many times, the result is a thick accumulation of P-rich rock such as the Phosphoria Formation, which certainly represents a unique sedimentary environment but not necessarily a unique, global P-cycling event. This is not to say that the global P cycle remains constant over time. On the contrary, global P cycling varied in the past, and below are two examples.
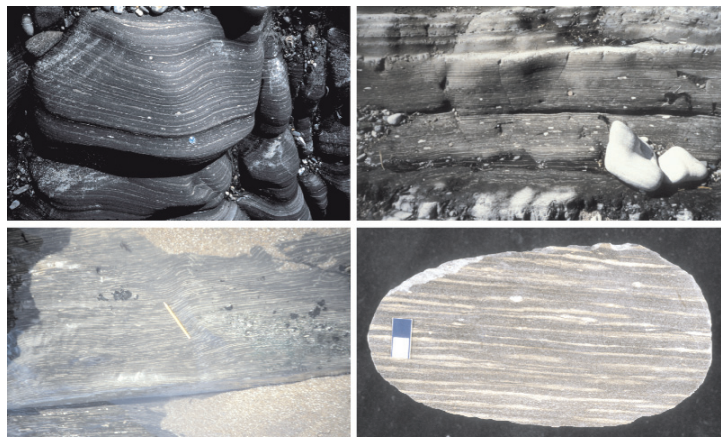
FIGURE 3 Phosphatic shales from the Miocene Monterey Formation, Shell Beach, central California, USA. The black shales are rich in organic matter, with up to 15 wt% total organic carbon (Filippelli 1997b), and are marked by whitish tan laminae and nodules of nearly pure carbonate-fluorapatite. These are seen clearly in the sliced section on the bottom right; scale is 2 cm. These phosphatic shales may be the result of minor reworking, but they are the raw materials for economically significant phosphorite deposits. Approximate height of top two images is 60 cm.

The Asian Monsoon, Weathering in the Himalayan-Tibetan Plateau, and the "Biogenic Bloom"

Studies of Cenozoic marine records have led to the interesting Himalayan uplift hypothesis (e.g. Raymo and Ruddiman 1992), which argues that the rise of the Himalayan-Tibetan Plateau dramatically increased net continental chemical weathering rates and drawdown of atmospheric CO_2 by weathering reactions; this resulted in global cooling and continental ice buildup through the Cenozoic. This hypothesis is largely based on the rapid rate of increase in the strontium isotope ratio of marine carbonates since 40 Ma, which has been cited (Raymo and Ruddiman 1992) as evidence for an increase in the rate of radiogenic (continental) strontium input to the ocean. The uplift and weathering of the Himalayan-Tibetan Plateau must have had a

profound effect on global ocean biogeochemical cycles; currently, dissolved chemical fluxes in rivers draining this region account for about 25% of the global total, although the region represents only slightly more than 4% of the global drainage area (Raymo and Ruddiman 1992). This region is thus a modern "weathering engine," with a combination of rapid physical weathering from uplift and rapid chemical weathering from high precipitation (FIG. 4). This weathering engine has gained its modern effectiveness since about 8 Ma, when Himalayan uplift became so great that it diverted tropical jet streams, creating an intensified Indian-Asian monsoon system (Molnar et al. 1993).

Records of sediment input and clay composition from the northern Indian Ocean reveal clear weathering changes in the plateau at this time, and Ge/Si ratios in opaline silica (FIG. 5) as well as biogenic sedimentation rates indicate that increased dissolved-element fluxes from this weathering event had an ocean-wide effect (Filippelli 1997a). It is likely that weathering intensity also increased in the Andes and Amazon Basin in this interval due to Andean uplift. Runoff intensification remobilized detrital material in foreland basins and induced intense chemical weathering. This weathering event increased the net flux of P to the ocean, as demonstrated by a peak in phosphorus accumulation rates at this time (FIG. 5). The temporary increase in nutrient inputs to the ocean triggered increased productivity and organic carbon burial, which can be observed in the carbon isotope record (FIG. 5), resulting in a widespread "biogenic bloom" and perhaps contributing to carbon dioxide drawdown in advance of Pliocene-Pleistocene glaciation (Filippelli 1997a).

FIGURE 4 Some parts of Tibet are still experiencing rapid uplift and physical erosion, as seen in this photo taken near Tingri, China. Several episodes of rapid erosion have produced high-energy braided channels and a recent river terrace (covered by rice paddies) that has been downcut by the modern river (PHOTO BY G. FILIPPELLI).
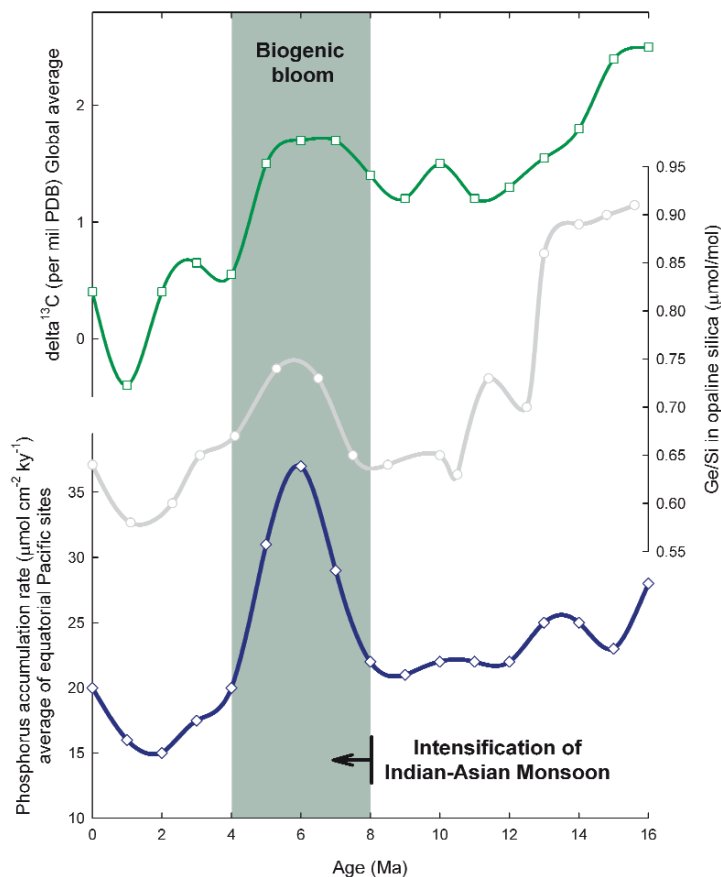

FIGURE 5 Geochemical record from marine sediments, showing dramatic transient shifts in P accumulation rate, Ge/Si in opaline silica, and carbon isotope composition (bulk marine carbonates) in the interval from just before 8 Ma to 4 Ma; this interval coincides with the intensification of the Asian monsoon (ORIGINAL DATA SOURCES IN FILIPPELLI 1997A).

Glacial–Interglacial Variations in the Oceanic P Cycle

What role does sea level change play in the marine P mass balance? This is an important question, given observed variations in shelf area caused by eustatic sea level changes on glacial timescales. The gain and loss of a sedimentary sink that is small in area but large in terms of net accumulation and biogeochemical dynamics certainly should play a role in whole-ocean geochemical budgets. The shelf-nutrient hypothesis (Broecker 1982) states that the loss in continental margin sinks for nutrients and carbon during glacial sea level lowstands (which resulted in a decrease in continental margin area of 60% during the last glacial maximum compared to now) should result in a net transfer of these components to the deep-ocean sink. The consequence would be significantly different and climatically mediated deep-ocean budgets for these elements. Many aspects of the shelf-nutrient hypothesis have been tested and debated, but little has been done to directly examine P budget changes in the ocean for this period.

This omission can be traced to several uncertainties in the P budget, but these aspects have now been clarified. First, new estimates of P residence time, based on a better understanding of deposition and regeneration, center around 20 ky (Colman and Holland 2000). This indicates that the P mass balance is responsive on glacial timescales. Second, the accumulation rates of biologically reactive P are now much more constrained, indicating that the continental margin sink, in spite of being less than 10% of the total ocean area, is almost the same as the deep-ocean sink in terms of accumulated P. This is a result of high sedimentation rates, lower recycling, and relatively high productivity at the margins (Filippelli 1997b). Third, it is plausible that material previously deposited on continental margins would be exposed to subaereal weathering during glacial lowstands. This is important because the CFA comprising the bulk of the reactive P in continental margin sediments is readily soluble under the weakly acidic conditions of subaereal exposure and weathering (Guidry and Mackenzie 2003). Finally, measurements of P export from the continents to the ocean on glacial timescales indicate high rates of P weathering and transfer during glacial terminations and rel-atively lower transfer rates during interglacial intervals (Filippelli et al. 2006); thus, the net increase in P delivered to the deep ocean during glacial periods would include not only that transferred from the margin sink but potentially also that derived from increased continental input.

Together, this new evidence highlights the potential impact that loss of the continental margin sink could have on oceanic P cycling, suggesting that this loss should be observable in deep-ocean records. In fact, we have recently demonstrated a profound change in the P mass balance at two widely separated areas: the southeastern Atlantic sector of the Southern Ocean and the eastern equatorial Pacific over the past 400 ky (Filippelli et al. 2007). Using a number of productivity proxies (including P accumulation rates, P/Ti as a reflection of "excess" P, and nanofossil accumulation rates), we observed broad peaks in enhanced productivity (FIG. 6). Each peak began during a glacial interval, reached a maximum just after the glacial–interglacial transition, and then decreased to a low value by the beginning of the next glacial interval. These records indicate that relatively high "excess" P export occurred about 40–60 ky after the onset of glacial intervals.

At face value, this result conflicts with the shelf-nutrient hypothesis given the lag in P response, but if the residence time of P in the ocean (10–20 ky) and mass balance laws are considered, this lag is predictable. We developed a simplistic model of the deep-sea response to addition of nutrients from the exposed continental shelf during lowered sea levels using sea level records (from oxygen isotope records of global ice volume) compared to the average P/Ti value of Southern Ocean and equatorial Pacific records (FIG. 6). The deep-sea nutrient model provides an extremely good fit to the composite P/Ti record with a 20 ky lag. Two remarkable aspects of these comparative records emerge from this analysis. First, even a simple comparison of excess P export results in a reasonable fit to a deep-sea nutrient model based on global sea level variations alone (FIG. 6). Second, the 20 ky lag, which provides the best long-term fit to the P/Ti data, is what would be expected given the P response time in the ocean as currently estimated (between 10 and 20 ky).
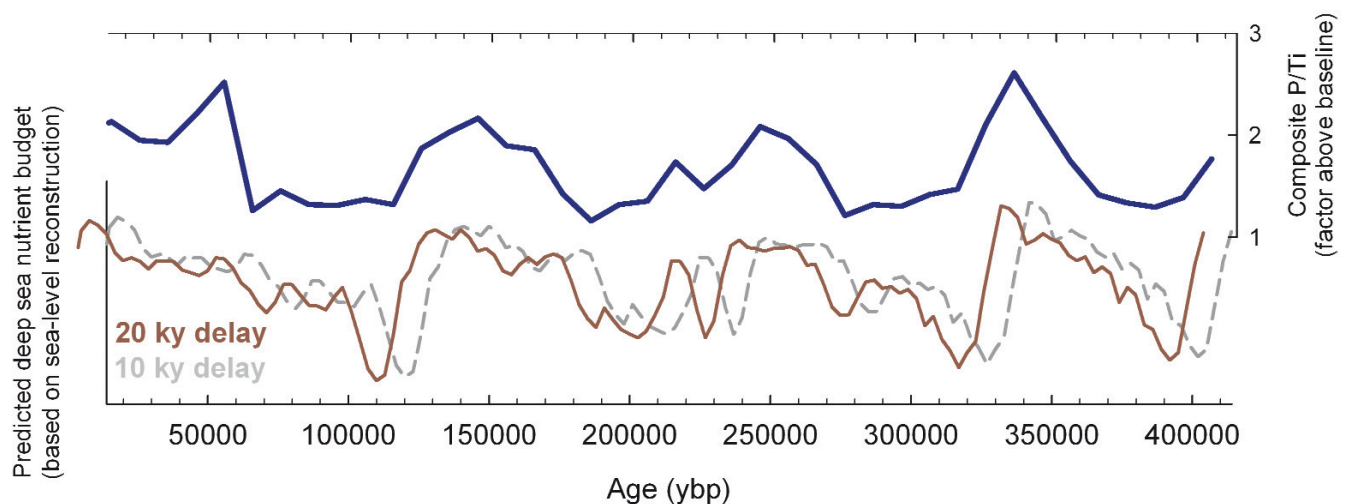

FIGURE 6 Changes in the mass balance of phosphorus on glacial timescales interpreted from the marine record of phosphorus burial (P/Ti), compared to a sea-level-based model (using the marine oxygen isotope record of ice volume) of phosphorus mass balance changes due to the redistribution of phosphorus from exposed continental margins during glacial sea level lowstands (after Filippelli et al. 2007). The P/Ti record is a composite of individual P/Ti records from the Southern Ocean and the equatorial Pacific Ocean (Filippelli et al. 2007). Note that the quantitative transfer of nutrients from continental margins recorded by the P/Ti record is best matched by the nutrient model with a 20 ky lag; thus the net impact of redistribution of marginal nutrients to the deep sea matches with the predictions of the P response time in the ocean.

The shifts in the oceanic P mass balance supported by these deep-ocean records have several important implications for global ocean productivity variations and carbon cycling. First, changing the depositional sink of P from the high sedimentation rate/short water column continental margins to the low sedimentation rate/deep water column deep-sea basins during glacial periods would enhance the degree of P recycling from particulate to dissolved material both in the water column and at the sediment–water interface, thus increasing the inventory of dissolved P in the oceans. Second, if the phosphate inventory is in fact increased during and just after glacial events, observed increases in surface-water productivity would not necessarily be the result solely of increased wind-driven upwelling. They could perhaps be caused by the same rate of water upwelling, but with an increased phosphate concentration. At this initial stage, the paleoceanographic P records provide deep-sea support for the "shelf-nutrient hypothesis" and should spur continued examination of geochemical mass balance variations through time.

THE GLOBAL PHOSPHORUS CYCLE: THE PRESENT AND FUTURE

The modern terrestrial P cycle is dominated by agriculture and human activity. Fertilizer-based food production and the application of P, along with N, K, and other micronutrients, in commercially available fertilizers boomed after World War II. This led to prosperity in some countries and exponential growth in global population. The use of fertilizer contributes substantially to the dissolved P cycle. Other contributors are deforestation, with its concurrent soil loss, and sewage and waste sources. Deforestation, typically by burning after selective tree harvesting, converts the standing stock of P in plant matter to ash. This P is rapidly dissolved, leached from the ash, and transported in rivers over timescales of a year or two (Schlesinger 1997). The lack of roots destabilizes the landscape, resulting in loss of the O and A soil horizons, relatively rich in organic P, from many of these areas. Some of the P from this soil is solubilized during transport. Human waste, waste from foodstuff processing, and the industrial use of detergents are additional,

although minor, contributors to anthropogenic P sources (FIG. 7A). Although detergents are now generally low in P, in the past they contained up to 7% by weight.

The human-era terrestrial P cycle, therefore, is substantially different from the prehuman cycle. This is demonstrated by the high loads of dissolved P in rivers, which are estimated at about twice natural values, and the higher loads of P-bearing particulates. Together, the net input of dissolved P from land to the oceans is 4–6 Tg P/y. This value represents a doubling of prehuman input fluxes, which clearly has led to eutrophication in coastal areas, and probably also has contributed to enhanced biological production in the whole ocean (FIG. 7B). One likely scenario is for the ocean to achieve a new steady state, in which the P output flux and the oceanic P reservoir will both increase in response to higher inputs. This new steady state would be short-lived, however, given the limited sources of P available for human exploitation and the relatively long residence time of P in the ocean. Nevertheless, a projection of anthropogenic increase to the ocean indicates that a sustained and significant eutrophication can be expected over the next two millennia (FIG. 7A). This projection does not predict global human extinction at 3600 AD (although it does not preclude it either); rather, the figure indicates the time at which projected global P resources for fertilizers will be depleted. Man will doubtless search for, and find, other avenues for recycling or concentrating P for fertilizer as the currently known resources become depleted, but these other sources will be much more limited in quantity and will not contribute "new" anthropogenic P to the system. Thus they do not impact a projection such as that shown in FIGURE 7A.

Eutrophication will impact the global carbon cycle, but will probably do little to offset anthropogenic carbon emissions. Based on the projection of P input to the ocean (FIG. 7A), the total excess input from 1600 to 3600 AD is 1860 Tg P. Given that, in the marine environment, between 106 and 170 units of C are buried per unit of P (Colman and Holland 2000), one can predict that excess phosphorus would effectively bury 76,000 to 123,000 Tg C. In essence, this burial
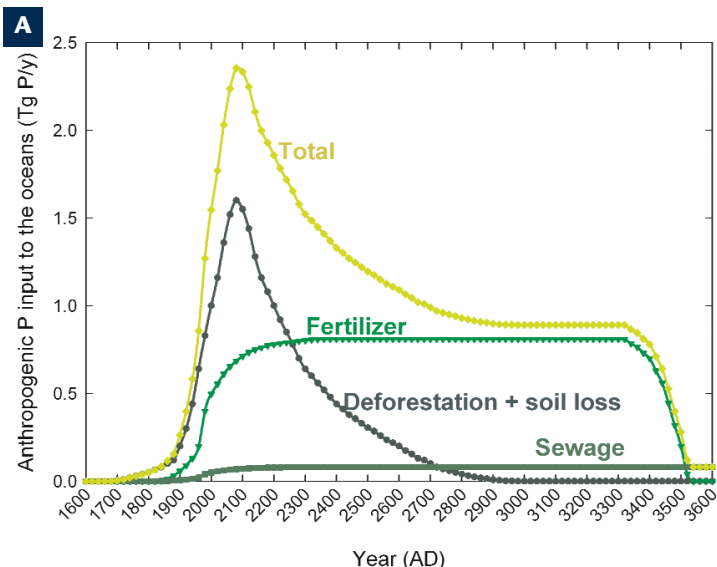
the depletion of known P reserves. The total integrated anthropogenic input of P (1600–3600 AD) is 1860 Tg. (**B**) A chalky, whitish blue bloom of marine algae called coccolithophoridae off Brittany, France; the bloom is the result of excess surface-water nutrients (eutrophication). Although coccoliths are small (micrometer scale), they often form large, concentrated blooms that are visible from space. PHOTO COURTESY OF NASA

FIGURE 7 (**A**) Reconstruction and projections of anthropogenic phosphorus delivered to the oceans as a result of several processes, including fertilization, deforestation and soil loss, and sewage. The projection is based on reports of population and arable land published by the World Health Organization and the Intergovernmental Panel on Climate Change. The fertilizer drop off coincides with

removes C from the atmosphere through the biological fixation of carbon dioxide during photosynthesis. The present annual rate of anthropogenic C addition to the atmosphere is 7900 Tg C (Marland et al. 2007), so the P eutrophication effect would only account for about 10–15 years of anthropogenic carbon emissions to the atmosphere over the next 2000 years (i.e. only ~0.6% of total projected carbon emissions, if emissions stay constant). Although the net effect as a carbon sequestration mechanism is minimal, the ecological impact of P fertilization to the ocean could be extreme. Given the other assaults on marine ecosystems, including warming and acidification of surface ocean waters from higher carbon dioxide levels, it would be pure speculation to project how P eutrophication would affect ecosystem structure and distribution in the future. However, those who have witnessed local eutrophication in ditches, streams, ponds, and lakes can attest to the ecological devastation that excess nutrients and the proliferation of monocultures can cause in such isolated environments. The eutrophication of coastal and open-marine ecosystems would result in a grim future for ecological diversity.

ACKNOWLEDGMENTS

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THE USE OF SUFFIXES IN MINERAL NAMES

The first hyphenated suffixes to be added to mineral names were used to distinguish polytypes: Frondel and Palache used these for wurtzite in 1950 and Smith and Yoder for micas in 1956. A joint nomenclature committee of the International Mineralogical Association and the International Union of Crystallography formalized these suffixes for polytypes, a modified Gard (1966) nomenclature, in a report published by Bailey in 1977.

Levinson had meanwhile introduced in 1966 a system of chemical-element suffixes for names of rare-earth minerals, which was revised and extended by Bayliss and Levinson in 1988. The system ("Levinson modifiers") is quite straightforward: a group name (without suffix) is given to a rare-earth mineral that has been identified without detailed chemical analysis of the rare-earth elements, e.g. monazite; a species name (with suffix) is given whenever the rare-earth element distribution has been determined, e.g. monazite-(Ce), monazite-(La), monazite-(Nd), and monazite-(Sm), the suffix indicating the dominant rare-earth element.

Such chemical-element suffixes have also been used in other mineral groups that can contain different substituting elements in one or more structural sites. Passaglia and Gottardi applied in 1973 these so-called "extended Levinson modifiers" to pumpellyite and julgoldite, and Moore in 1978 applied them to jahnsite and whiteite. The latter have suffixes consisting of three elements because solid solution occurs in three structural sites, e.g. jahnsite-(CaMnFe) and whiteite-(MnFeMg).

An earlier (1971) proposed use of an extended Levinson modifier for the naming of the Na-dominant analogue of zippeite had met with heated discussion within the IMA Commission on New Minerals and Mineral Names (now the Commission on New Minerals, Nomenclature and Classification, CNMNC). The choice was whether to name the mineral sodium-zippeite or zippeite-(Na). One member even offered a Solomon-like solution for the problem: write the name as zippeite-(Na) but pronounce it as sodium-zippeite. At that time, the Commission decided on sodium-zippeite, in writing and in word!

Chemical-element suffixes have recently been applied in the revision of nomenclature schemes for several mineral groups: zeolites (Coombs et al. 1997), labuntsovites (Chukanov et al. 2002), arrojadites (Chopin et al. 2006) and epidotes (Armbruster et al. 2006). The suffixes of zeolites (e.g. stilbite-Ca and stilbite-Na) and labuntsovites (e.g. tsepinite-K and tsepinite-Sr) are used without parentheses to indicate extra-framework cations. Application of the suffix-based nomenclature is not compulsory in the revision of mineral groups. As always with nomenclature decisions, the CNMNC adheres to its traditional principle that each case should be considered on its own merits. The many cation sites in minerals of the eudialyte group would cause names like kentbrooksite-(Fe^{3+}SiCaOH), and the authors of the report (Johnsen et al. 2003) thus decided on a unique-name system for this group, the example given here being "feklichevite."

Nomenclature revisions of mineral groups sometimes involve the unavoidable disappearance of well-known mineral names, a result not always met with agreement in the mineralogical community: the 2006 renaming of hancockite into epidote-(Pb) was greeted with considerable acrimony and was called "an eternal insult from the IMA" in one editorial in a collector publication.

Certainly, the CNMNC recognizes the value of the amateur mineralogical community and takes seriously the comments voiced in various forums. For many years there have been complaints over the naming of members of several common mineral series (e.g. apatite and columbite) for which chemical analyses (or other tests) are needed to distinguish the correct identity (and name) of the series member. To address this, the CNMNC decided in September 2007 to change the names in a number of mineral groups to suffix-based names, namely apatites, ellestadites, apophyllites, axinites, columbites, tantalites, tapiolites and pyrosmalites. The occasion was also used to tidy up other categories of mineral names that, in fact, did not follow the Commission's own rules:

two-word names, names with superfluous hyphens, and names with incorrect diacritical marks. The complete lists haven been published in the *Mineralogical Record*, volume 39 (2008), pages 131–135. Therein it is also stated that the vast majority of the about 500 existing mineral names with prefixes that indicate a structural analogue or polymorph should remain unchanged.

But even before that publication appeared in print, a new problem arose in the apatite group. Because of the existence of a Sr-dominant apatite, previously named "strontium apatite," the new suffix names for the apatites include Ca or Sr in addition to the dominant anion. For example, "hydroxylapatite" is now apatite-(CaOH) and "strontium apatite" is now apatite-(SrOH). However, the recent discovery of an even more Sr-dominant apatite spoiled the game and made it clear that things in the apatite group are a bit more complex. There is indeed not a simple solid-solution series between pure Ca apatite and pure Sr apatite. Intermediate members, such as the existing "strontium apatite" and fluorcaphite (in the F-dominant series), have cation ordering, comparable to the cation ordering in the intermediate-member dolomite in the calcite–magnesite series.

In March 2008 a subcommittee was formed to resolve the nomenclature problems in the entire apatite group, including phosphates, arsenates, vanadates and silicates with the apatite structure. In light of that, perhaps it would be wise to wait some time before changing apatite-group names on specimen labels and in catalogues! Life is not simple, and neither is mineralogy, apparently.
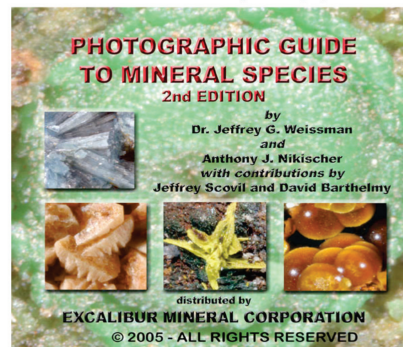
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Many of the cited literature references are available on the CNMNC website:
www.geo.vu.nl/~ima-cnmmn/imareport.htm