Rise of angiosperms as a factor in long-term climatic cooling

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ABSTRACT

By Late Cretaceous or early Tertiary time, the diversification and proliferation of angiosperm-deciduous ecosystems resulted in higher rates of mineral weathering. This increase in the global average weathering rate would have caused a decrease in atmospheric carbon dioxide and, hence, global cooling. The magnitude of this decrease is quantitatively explored here by developing a formulation for global weathering which combines ecosystems that differ in their fractional global coverage and intrinsic rates of weathering. Incorporating this formulation into models—specifically, several previously developed global steady-state models of the geochemical cycle of carbon between the atmosphere and carbonate rocks—gives results that show significant global cooling from the evolution of the angiosperm-deciduous ecosystems. This cooling may vary from a few degrees up to 10 °C. In this way, deciduous ecosystems with high rates of mineral weathering could have contributed to the evolution during the past 100 m.y. of a cooler Earth and thus were a factor in producing conditions that enhanced their global proliferation.

INTRODUCTION

Knoll and James (1987) showed that different ecosystems cause weathering of soil minerals at different rates. Specifically, following Likens et al. (1977), they indicated that angiosperm-deciduous ecosystems lose soil K, Mg, and Ca at rates three to four times higher than do conifer-evergreen ecosystems. The number of genera and species of angiosperms grew dramatically between about 120 and 80 Ma (Lidgard and Crane, 1988), and serious competition with conifers occurred by Late Cretaceous to early Tertiary time (Knoll and James, 1987). Knoll and James pointed to the origin and diversification of grasses during the Cenozoic as another angiosperm-linked increase in weathering rates.

The fluxes of Ca and Mg ions from continental silicates to the oceans where they are precipitated in carbonates is a crucial part of the global carbonate-silicate cycle on geologic time scales (Walker et al., 1981; Berner et al., 1983). Associated with the transfer of these cations from the silicate to carbonate rock reservoirs is the transfer of carbon dioxide from the fluid ocean-atmosphere reservoir to carbonate rocks; the carbon dioxide is returned to the atmosphere by the reverse process, the metamorphism and magmatic conversion of carbonates into silicates. On time scales roughly greater than 10^5 yr, the two processes are in balance: the removal of carbon dioxide from the atmosphere by the weathering of silicates equals its release to the atmosphere by tectonic activity. Quantifying our understanding of the specific dynamics of this balance, which determines the amount of carbon dioxide in the atmosphere, requires mathematical models (Walker et al., 1981; Berner et al., 1983; Volk, 1987).

Because evolutionary changes in the life strategies of vegetation affect the rates of weathering, they should also affect Earth's atmospheric carbon dioxide level and climate. Here, we use the methods of the geochemical-cycle models to estimate the effects on climate of changes in the weathering rates that are rooted in plant ecology.

ECOLOGICAL TERM FOR THE CARBONATE-SILICATE GEOCHEMICAL CYCLE

Models of the carbonate-silicate cycle have developed weathering parameters as an explicit function of the following: global average temperature (T) (Berner et al., 1983); T and atmospheric p_{CO_2} (P_{atm}) (Walker et al., 1981); and P_{atm}, T, and soil p_{CO_2} (P_{soil}) (Volk, 1987). The problem here is to provide a way to incorporate the biological effects on weathering into these models. These past studies dealt with terrestrial biology with varying degrees of explicitness, but none allowed for the possibility that different large-scale terrestrial ecosystems, with different weathering rates, could coexist in the same environment (P_{atm} and T).

To quantify this, we can infer a world with two types of ecosystems, one with a high weathering rate per unit area (W_H) and one with a low rate (W_L). The two can be related by an ecological weathering-rate ratio (W_{ratio}), defined by

$$W_{ratio} = \frac{W_H}{W_L}. \tag{1a}$$

The two ecosystems cover different fractions of the continents, and for convenience the sum of the fractional areas, A_H and A_L, is assumed to be constant and equal to unity:

$$A_L + A_H = 1. \tag{1b}$$

The simplification of Earth's ecology into this binary system is in the spirit of developing parameters for sets of complex global processes to examine the consequences of their coupling (Berner et al., 1983). The high and low weathering cases would correspond to the distinction between angiosperm-deciduous and conifer-evergreen eocsystems, discussed in a global and evolutionary framework by Knoll and James (1987). In this case, an average global ecological weathering rate (W_E) can be defined:

$$W_E = A_H W_H + A_L W_L. \tag{1c}$$

The W_E term could be incorporated into any of the previously developed models of the carbonate-silicate cycle by multiplying the original weathering-rate term in a given model by a new term that is a function of W_E. Because the parameters of the geochemical models were developed by calibrating to various fluxes in today's world, it is necessary to incorporate the new W_e term so as not to affect the parameters and present fluxes of a given model. This is accomplished by defining a normalized global ecological weathering rate (E) as

$$E = \frac{W_E}{W_{E,o}}. \tag{1d}$$

Note that E is global and can vary with time; also, E is conceptually distinct from W_{ratio}. Since by this definition $E_o = 1$, multiplying any model's original weathering rate by E will not change the calibration for today's world, but will allow us to examine the effects of changing the relative areas to simulate a past state by varying A_H and A_L while W_H and W_L remain constant. Specifically, following the lines of reasoning in Knoll and James (1987), we assume the existence of some point in the past prior to the evolution of the high-weathering ecosystems. When $A_{H,past} = 0$ and $A_{L,past} = 1$, from equation 1c, $W_{E,past} = W_L$. By using equation 1c to define $W_{E,o}$ (knowing $A_{H,o}$ and $A_{L,o}$), equations 1a–1d can be combined to calculate E_{past} in terms of $A_{H,o}$ and W_{ratio}:

$$E_{past} = \frac{1}{1 + A_{H,o}\,(W_{ratio} - 1)}\,. \tag{1e}$$

Note that if W_{ratio} is greater than 1 and $A_{H,o}$ is greater than 0, E_{past} will be less than E_o.

TWO SIMPLE GEOCHEMICAL-CYCLE MODELS

With equations 1a–1e, we look first at the model of Berner et al. (1983). This model was simplified into a steady-state solution valid for times greater than about 10^5 yr by Kasting (1984); the changes in P_{atm} and T over the past 100 m.y. driven by changes in global sea-floor generation rate and continental area in the full, more complex model of Berner et al. are reproduced closely by the simpler solution (Volk, 1987). Here, we incorporate the term E (E_o or E_{past}) in the right-hand side of the equation for the global total weathering rate, f_{WR}, and the simple, steady-state system becomes

$$f_{WR} = E\,(1 + 0.038\,[T - T_o])\,(1 + 0.049\,[T - T_o]), \tag{2a}$$

$$T - T_o = 2.88\,\ln\left(\frac{P_{atm}}{P_{atm,o}}\right). \tag{2b}$$

Inside the first parentheses on the right-hand side of equation 2a is the global continental runoff factor; inside the second parentheses is the factor for global average bicarbonate-ion concentration in river runoff. Both factors were developed as functions of global temperature, T, and the present temperature, T_o (Berner et al., 1983). Constants have dimensions that make f_{WR} nondimensional ($f_{WR,o} = 1$); E is also nondimensional. Equation 2b is for the CO_2-greenhouse relation.

A crucial point is that over long time periods, if the geophysical forcings of tectonic generation of CO_2 and change in continental area are taken as constant (to isolate the ecological factor in this study), the global total weathering rate, f_{WR}, will be constant and equal to unity (nondimensional, normalized to today). This is true despite the changes in weathering rate brought about by the evolution of the ecosystem with the high-weathering rate. The reason for this is as follows: an "instantaneous" evolution and spread of such an ecosystem would increase the global weathering rate for a given environment of P_{atm} and T. However, for a constant supply of CO_2 to the ocean-atmosphere system from metamorphic and magmatic decarbonation and constant total continental area, both P_{atm} and T would gradually decrease in a transition to restore the global weathering rate, f_{WR}, again to unity. The new world would have a higher ecology-caused weathering rate, balanced by lower temperature and p_{CO_2}-caused weathering rates related to the decreased P_{atm} and cooler climate.

Walker et al. (1981) separately developed an alternative system for the geochemical carbon cycle that differs from that of Berner et al. (1983), yet is similar in many of the general quantities expressed. Volk (1987) extended this system to include the effects of changes in P_{atm} upon average global terrestrial productivity (Π) and, in turn, soil pCO_2 (P_{soil}). Other terms include the minimum P_{atm} necessary for net photosynthesis (P_{min}), a maximum value of Π (Π_{max}), and the value of P_{atm} at which $\Pi = 0.5\,\Pi_{max}$ ($P_{½}$). The complete system is

$$f_{WR} = E\left(\frac{P_{soil}}{P_{soil,o}}\right)^{0.3}\exp\left(\frac{T - T_o}{13.7}\right), \tag{3a}$$

$$\frac{P_{soil}}{P_{soil,o}} = \frac{\Pi}{\Pi_o}\left(1 - \frac{P_{atm,o}}{P_{soil,o}}\right) + \frac{P_{atm}}{P_{soil,o}}, \tag{3b}$$

$$\Pi = \Pi_{max}\,\frac{P_{atm} - P_{min}}{P_{½} + (P_{atm} - P_{min})}, \tag{3c}$$

$$P_{½} = \left(\frac{\Pi_{max}}{\Pi_o} - 1\right)(P_{atm,o} - P_{min}), \tag{3d}$$

$$T - T_o = 4.6\left(\frac{P_{atm}}{P_{atm,o}}\right)^{0.364} - 4.6. \tag{3e}$$

Here, the global weathering rate is an explicit function of T and P_{soil} (where the mineral weathering occurs). Typically 10 to 100 times higher than P_{atm} (Holland et al., 1986), and derived by balancing the CO_2 outgassed by the biota into the soil and its escape by diffusion to the atmosphere (Volk, 1987), P_{soil} in equation 3b is a function of P_{atm} and Π. The formulation of Π in equation 3c represents a combination of the effects of quantum photosynthetic yield, water-utilization efficiency, continental water supply, changes in the growing season and arable land, and a transpiration-rainfall feedback (Volk, 1987). Equation 3c is a Michaelis-Menton relation, nearly linear for small values of P_{atm}, whereas at higher values for P_{atm} it approaches the maximum value, Π_{max}. The equation 3d for $P_{½}$ is derived by calibrating to present conditions (all subscripts o) and P_{min} is typically 0.2 P_{atm}. Equation 3e is the CO_2-greenhouse formulation from Walker et al. (1981), approximately equivalent to equation 2c. $P_{soil,o}$ will be taken as 30 $P_{atm,o}$; this is a reasonable mid-value from the range given in Holland et al. (1986). Here, the system is extended by the additional term E in the weathering equation 3a, which is analogous to adding E in equation 2a.

RESULTS OF THE MODELS

Equations 2a–2b and 3a–3e constitute two separate model systems (hereafter termed models 1 and 2, respectively) for examining and isolating possible effects on climate from changes in weathering caused by plant ecology. Note that the normalized global ecological weathering rate, E, is conceptually formulated to be independent of productivity changes related directly to P_{atm}. This follows Knoll and James (1987) in attributing the differences in weathering rates of angiosperm-deciduous and conifer-evergreen ecosystems to specific structural differences in their life cycles in space and time, not to differences in annually averaged productivity (although productivity differences are certainly possible).

The ecological weathering-rate ratio of equation 1a, W_{ratio}, is taken as an independent parameter, and quantifying its effect on global temperature is a major goal of this study. It is varied between 1 and 4, because 4 is the enhancement of K loss by the angiosperm-deciduous ecosystems and is assumed to be a maximum possible value for the enhancement of soil loss of the Ca and Mg cations, essential in the cycle of carbon between the silicate and carbonate rock reservoirs. The enhancement of loss of Ca and Mg cations was cited as about a factor of 3 in Knoll and James (1987).

To explore the sensitivity of the results to the global-fractional area of the ecosystems with high weathering rates, both models are run with two discrete cases: $A_{H,o} = 0.1$, and $A_{H,o} = 0.5$. The calculation proceeds as follows: With $E = E_o = 1$, all variables in the two models are equal to their present values, $T = T_o$; $P_{atm} = P_{atm,o}$; $P_{soil} = P_{soil,o}$; and $\Pi = \Pi_o$. Then, maintaining for the steady-state $f_{WR} = 1$, as discussed above, and setting the parameters $A_{H,o}$ and W_{ratio}, E_{past} is calculated from equation 1e. In that E_{past} is less than E_o, results for both models yield higher values for T and P_{atm} in this past state. These higher values represent the environmental conditions that could have changed to their present values by the origin and spread of the ecosystems with high weathering rates.

Figure 1a and 1b shows the results for T and P_{atm}. To facilitate the interpretation, Figure 1a is plotted as $T_o - T_{past}$ and should be read as the cooling of Earth due to the contribution to increased weathering rates, virtually synchronous with the spread of the high-weathering ecosystems.

When $A_{H,o} = 0.5$ and $W_{ratio} = 3$ (the value cited by Knoll and James), the two models calculate coolings of about 6 to 10 °C. As apparent in Figure 1b, such cooling corresponds to a reduction in P_{atm} to between 3% and 10% of its past value.

Even in a world with $A_{H,o} = 0.1$, the reduction in P_{atm} and T caused by the high-weathering ecosystems can be significant; about 50% and 1 to 2 °C. For any given $A_{H,o}$ and W_{ratio}, model 2 gives less cooling and less reduction in P_{atm} than does model 1. The explicit incorporation in model 2 of the CO_2 enhancement of productivity accounts for about one-third of the difference, and the remaining two-thirds result from different parameterizing of the physical effects of soil pCO_2 and T on weathering.

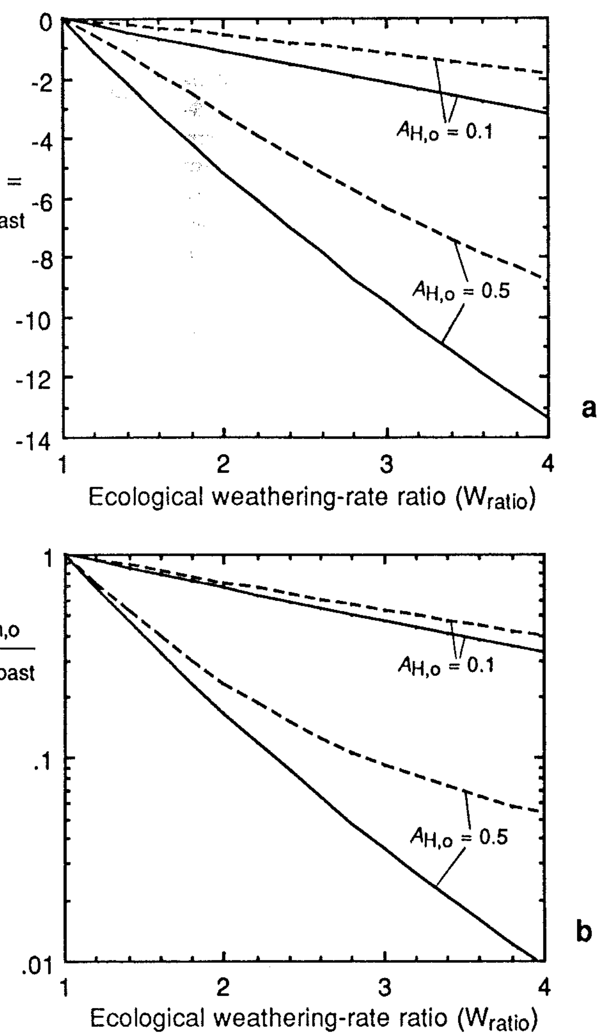

Figure 1. a: Global average surface temperature as function of W_{ratio} and $A_{H,o}$ using two independent geochemical models. W_{ratio} is ratio of weathering rates from ecosystems with high rates of weathering per unit area to those with low rates. $A_{H,o}$ is fractional area of ecosystems with high weathering rates for "present" Earth. Temperature is plotted as T_o–T_{past} and represents amount of cooling from some point in past caused by evolution and global spread of (angiosperm) ecosystems with high weathering rates. Solid lines are from model 1 (equations 2a and 2b); dotted lines are from model 2 (equations 3a–3e). Both models additionally use equation 1e. b: Atmospheric CO_2 (P_{atm}) as function of W_{ratio} for same conditions in a. P_{atm} and T are related through CO_2-greenhouse equations (2b and 3e for two models, respectively). Graph is plotted as $P_{atm,o}/P_{atm,past}$ and represents relative change in P_{atm} from past Earth with only ecosystems with low weathering rates to Earth with ecosystems with both low and high weathering rates.

DISCUSSION

What fraction of Earth is covered by high-weathering ecosystems? We can examine the data on ecosystem types in Atjay et al. (1979, see Table 5.2, which quotes Whittaker and Likens, 1975) for systems that are predominantly angiosperms with strong seasonal cycles, characteristics broadly discussed by Knoll and James (1987) as giving high weathering rates. If we discount the areas of the very low productivities of the desert and semidesert scrub (their possible increased productivity in a world with higher T and P_{atm} is already accounted for in the formulation for Π in equation 3c; see Volk, 1987), the global terrestrial area is 105×10^{12} m^2. Temperate deciduous forests with areas of 7×10^{12} m^2 and temperate grasslands with areas of 9×10^{12} m^2 are apparently included in the characteristics outlined by Knoll and James. Certainly a fraction of the cultivated land (14×10^{12} m^2) should be included (either in its present or former state), some of the woodland and shrubland (8.5×10^{12} m^2), and possibly some of the tropical seasonal forest (7.5×10^{12} m^2) and savanna (15×10^{12} m^2). Together, the temperate deciduous forests and the temperate grasslands are 15% of the 105×10^{12} m^2 total. If we assume that 50% of each of the other four categories should be included, the yield is 36% of the 105×10^{12} m^2. It therefore seems reasonable that the calculation with values for $A_{H,o}$ of between 0.1 and 0.5 broadly brackets the actual area ranges for the high-weathering ecosystems.

Knoll and James (1987) reasoned that the higher weathering rates of the angiosperm-deciduous systems were a result of the discontinuous nature of leaf loss with the related enhanced chance of severe ion loss in the runoff, in comparison to the conifer-evergreens, which have more efficient cycling of nutrients, gradual ion addition to the soil, and longer nutrient immobilization times in the leaves. The data of Likens et al. (1977) used to quantify these differences for the Ca ion loss—which is between two and three times higher than the Mg ion loss and therefore of primary importance in the geochemical carbon cycle—included nine angiosperm-deciduous ecosystems and ten conifer-evergreen systems. These data show that the loss rates of Ca ion for the angiosperm-deciduous and conifer-evergreen systems in kg/ha/yr are, respectively, 11.7 vs. 6.9 for the medians ($W_{ratio} = 1.7$); 39.9 vs. 11.2 for the means with all points ($W_{ratio} = 3.5$); 27.4 vs. 7.3 for the means with highest and lowest values for each system excluded ($W_{ratio} = 3.8$); and 11.5 vs. 3.7 for the means with two highest and two lowest values excluded for each system ($W_{ratio} = 1.7$). Thus, even though there is a large spread of values in the individual locations of the two ecosystems, they support the range of values for W_{ratio} explored in Figure 1.

The formulation for E, the normalized global ecological weathering rate, introduced into the geochemical models in this work was developed to operate in addition to weathering factors from direct effects of atmospheric CO_2 on terrestrial productivity. For example, here $P_{soil,o}$ does not differ for the two ecosystem types. But ecological differences that affect weathering through P_{soil} could exist. For example, peak seasonal P_{soil} values for wheat, corn, and soybeans (5%–7%; Buyanovsky and Wagner, 1983) are about an order of magnitude higher than those of a spruce-fir forest (0.6%–0.8%; Fernandez and Kosian, 1987).

CONCLUSIONS

Crowley (1983) reviewed the substantial amount of evidence—for example, from oxygen isotopes of benthic carbonates, absence or presence of continental ice, and paleobotany—that indicates general global cooling trends over the past 100 m.y. A strong case for linking this cooling to a decrease in the rate of sea-floor generation, which decreases the metamorphic and magmatic source of CO_2 and therefore decreases the steady-state P_{atm} and T, was a major result of a geochemical modeling study by Berner et al. (1983). The potential climatic effect of the evolution and spread of ecosystems with high weathering rates shown here is comparable to the geophysical climatic effects in the Berner et al. study.

It is possible that the evolution of angiosperms and deciduousness operated as part of a positive feedback loop. If the evolution of ecosystems with high weathering rates cooled Earth, they may have helped to create the conditions—i.e., strong seasonality—that enhanced their success in competition with the conifer-evergreen systems. This might serve as a specific mechanism for the proposal by Lovelock and Whitfield (1982) that the terrestrial biosphere would reduce P_{atm} over time by changes in the weathering process. A system that used deciduousness and/or increased production of organic soil acids, as discussed for land plants in general by Knoll and James (1987), to mobilize nutrients would gradually alter the climate to produce the climatic regime that possibly gave it an advantage over the evergreen systems, particularly in certain latitude belts on Earth. Changes in the amplitude of the seasonal temperatures due to shifts in the positions of continents, demonstrated by Crowley et al. (1986), would result in lower global average temperatures if the area of the ecosystems with high weathering rates increased. However, the nature of these suggested couplings could be quite complex. For example, it is intriguing to speculate that the presence of boreal forests with low weathering rates in the very high latitudes might limit the magnitude of a cooling trend by becoming more widespread as their particular climatic regime grows with deep cooling.

Significant cooling would have resulted from the evolutionary origin and spread of ecosystems with high rates of mineral weathering. The potential importance of life in the evolution of climate shown here demands that work should continue to combine geochemical models with ecological data.

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Reviewer's comment

This paper is an outstanding addition to the theoretical modeling of atmospheric evolution.
