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The rise and fall of continental arcs: Interplays between magmatism, uplift, weathering, and climate

Cin-Ty A. Lee [a,*], Sally Thurner [a], Scott Paterson [b], Wenrong Cao [b]

[a] Dept. Earth Science, Rice University, MS-126, 6100 Main St., Houston, TX 77005, United States
[b] Dept. Earth Sciences, University of Southern California, 3651 Trousdale Pkwy, Los Angeles, CA 90089-0740, United States

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A B S T R A C T

Continental arcs, such as the modern Andes or the Cretaceous Sierra Nevada batholith, are some of the highest topographic features on Earth. Continental arc volcanoes may produce more CO_2 than most other types of volcanoes due to the interaction of magmas with sedimentary carbonates stored in the continental upper plate. As such, global flare-ups in continental arc magmatism may amplify atmospheric CO_2 concentrations, leading to climatic warming. However, the high elevations of continental arcs may also enhance orographic precipitation and change global atmospheric circulation patterns, possibly increasing the efficiency of chemical weathering and drawdown of atmospheric CO_2, which may subdue the climatic warming response to volcanic activity. To better evaluate the climatic response, we develop models that integrate magmatic crustal thickening, topographic uplift, isostasy and erosion. The topographic response is used to predict how soil formation rates, soil residence times, and chemical weathering rates vary during and after a magmatic episode. Although magmatism leads to crustal thickening, which requires topographic uplift, highest elevations peak ~10 My after magmatism ends. Relatively high elevations, which enhance erosion and chemical weathering of the continental arc, persist for tens of million years after magmatism ends, depending on erosion kinetics. It has recently been suggested that the Cretaceous–Paleogene greenhouse (high atmospheric CO_2 and warm climate) coincided with a global chain of continental arcs, whereas mid- to late Cenozoic icehouse conditions (low atmospheric CO_2 and cold climate) coincided with a lull in continental arc activity after 50 Ma. Application of our models to the Sierra Nevada (California, USA) continental arc, which represents a segment of this global Cretaceous–Paleogene continental arc, reproduces the observed topographic and erosional response. Our models require that the newly formed continental arc crust remained high and continued to erode and weather well after (>50 My) the end of magmatism. Thus, in the aftermath of a global continental arc flare-up, both the total volcanic inputs of CO_2 decline and the average weatherability of continents increases, the latter due to the increased proportion of widespread remnant topography available for weathering and erosion. This combination leads to a decrease in the long-term baseline of carbon in the ocean/atmosphere system, leading to cooling. Mid-Cenozoic cooling is often attributed solely to increased weathering rates associated with India–Eurasian collision and the Himalayan orogeny. However, the total area of now-extinct Cretaceous–Paleogene continental arcs is 1.3–2 times larger than that of the Himalayan range front and the Tibetan plateau combined, suggesting that weathering of these remnant volcanic arcs may also play a role in drawing down CO_2 through silicate weathering and subsequent carbonate burial. In summary, if global continental arc flare-ups lead to greenhouse conditions, long-lived icehouse conditions should follow in the aftermath due to decreased CO_2 inputs and an increase in regional weathering efficiency of remnant arc topography.

1. Introduction

Climate, that is, Earth's surface temperature, is modulated by solar insolation, albedo and greenhouse gas concentrations, such as CO_2, in the atmosphere. Of interest here are long-term (>10 My) variations in Earth's climatic baseline. For example, Earth appears to vacillate between long (>50 My) warm intervals, where there are no significant continental ice sheets like the Cretaceous–Paleogene, to long cold intervals, where ice sheets are important, such as the mid-Cenozoic to present (Zachos et al., 2008). Compared to the Archean, when the sun was dimmer and there may

* Corresponding author.
E-mail address: ctlee@rice.edu (C.-T.A. Lee).

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Nomenclature

h	elevation	km
H	excess thickness relative to continental crust at sea level	km
H_{tot}	total thickness of continental crust	km
h_b	baseline thickness of continental crust at sea level	km
h_{ec}	depth to garnet pyroxenite (65 km)	km
H_{ec}	thickness of eclogite layer	km
$H_{ec,max}$	maximum thickness of eclogite layer before foundering	km
ρ_c	depth-averaged density of continental crust (2.87 g/cm^3)	g/cm^3
ρ_{ec}	density of garnet pyroxenite (3.5 g/cm^3)	g/cm^3
ρ_m	density of mantle peridotite (3.33 g/cm^3)	g/cm^3
t	time (My = millions of years; Ma = millions of years ago)	
M	magmatic flux (km^3 km^{-2} My^{-1})	km/My
M_{max}	peak magmatic flux	km/My
E	erosion rate	km/My
β	erosional constant	km/y
w	width of active arc	km
p	precipitation rate	mm/y
$\tau_E = w/\beta$	erosional response time	My
$\tau_E = (k_e p)^{-1}$	erosional response time	My
$\tau_L \equiv \tau_E/(1 - \rho_c/\rho_m)$	landform response time	My
$H_\infty = M\tau_L$	steady state excess thickness for constant $\dot{M}$	km
$h_\infty = M\tau_E$	steady state elevation for constant $\dot{M}$	km
z	soil mantle thickness	m
P	soil production rate	km/My
P_o	soil production rate when z = 0	km/My
D	e-fold decay lengthscale of soil production function P	m
$\tau_S = D/E$	soil formation response time	My
$z_\infty = D\ln(P_o/E)$	steady state soil mantle thickness	m
τ_{res}	soil residence time	My
k_w	kinetic rate constant for chemical weathering (inverse response time)	s^{-1}
$f_w = (1 + k_w\tau_{res})^{-1}$	fraction of weatherable material remaining in soil	
$1 - f_w$	fraction of material lost by chemical weathering	

have been fewer continents, long-term variations in solar insolation and albedo (mostly due to exposed continental surface area) during the Phanerozoic were not large enough to play an important role in long-term variation in climate (Caldeira and Kasting, 1992). This leaves variations in atmospheric CO_2 as the primary influence on the long-term climatic baseline.

On >1 My timescales, the fluxes of carbon within the exogenic system, that is between the ocean, atmosphere and biosphere, are closely balanced, which means that long-term variations in atmospheric CO_2 are controlled by inputs from the endogenic to the exogenic system via volcanism, metamorphism, and oxidative weathering of fossil organic carbon and by outputs from the exogenic system via silicate weathering and biological productivity in the form of carbonate and possibly organic carbon burial (Berner, 1991; Berner et al., 1983; Kerrick, 2001; Kerrick and Caldeira, 1994; Ridgwell and Zeebe, 2005; Walker et al., 1981). Silicate weathering rates scale with atmospheric CO_2, hence carbonate precipitation represents a negative feedback (Walker et al., 1981). To illustrate, the rate of change of carbon in the ocean/atmosphere system on long timescales is given by

$$\frac{dM_{oa}}{dt} = J_{in} - kM_{oa} \tag{1}$$

where M_{oa} is the mass of carbon in the ocean + atmosphere system, J_{in} represents volcanic and metamorphic degassing, and the second term represents carbonate deposition in response to silicate weathering with k being a kinetic rate constant describing the global efficiency of weathering (e.g., weatherability). In reality, weathering rates (e.g., kM_{oa}) are a complex function of a wide variety of factors, so the linear kinetics assumed here are meant only to illustrate a concept. Thus, in Eq. (1), the inverse of k (e.g., $1/k$) is the response time of the ocean + atmosphere system to perturbations. Because the response times for silicate weathering and subsequent carbonate burial are less than 100 ky, inputs and outputs should be in balance and the global C system should be near steady state on timescales >1 My (Berner et al., 1983; Zeebe and Caldeira, 2008). Global CO_2 drawdown via silicate weathering can only be as large as global CO_2 inputs J_{in}.

Climate, however, clearly varies on timescales greater than 1 My, as evidenced by long-lived greenhouse and icehouse oscillations throughout Earth's history. What drives these long term secular changes is one of the most perplexing problems in the Earth sciences. Since the left side of Eq. (1) is zero on long timescales, the long-term steady state carbon content in the ocean + atmosphere is given by

$$M_{oa} \sim \frac{J_{in}}{k} \tag{2}$$

Thus, while global CO_2 drawdown and CO_2 inputs are balanced, long term changes in the actual carbon content of the atmosphere + ocean system are possible if there are changes in the external inputs driving the system (J_{in}) or in the overall weathering kinetics (k) modulating the system or both. For example, secular variations in the dynamics of the Earth's deep interior might set CO_2 inputs (Kerrick, 2001; Kidder and Worsley, 2010; Larson, 1991). Increased mid-oceanic ridge spreading and eruption of flood basalts have been suggested to have driven the long-lived (150–50 Ma) Cretaceous–Paleogene greenhouse (Berner, 1991; Hong and Lee, 2012; Kidder and Worsley, 2010; Larson, 1991). Mid-Cenozoic cooling has been suggested to be linked to India–Eurasia collision and the rise of the Tibetan plateau, which led to enhanced kinetics of physical/chemical weathering (Edmond, 1992; France-Lanord and Derry, 1997; Raymo and Ruddiman, 1992; Zachos and Kump, 2005). Thus far, most studies treat changes in inputs and weatherability independently.

What happens when magmatism and mountain building are linked? Such may be the case for volcanic arcs. Volcanic arcs are a large source of CO_2, which derives from background mantle CO_2, decarbonation of subducted carbonates, and magmatically induced thermal metamorphism of carbonates in the upper plate (Burton et al., 2013; Dasgupta and Hirschmann, 2010; Hilton et al., 2002; Lee and Lackey, 2015; Lee et al., 2013; Marty and Tolstikhin, 1998). Averaged over 10–50 My, intra-oceanic arcs may not be as important as continental arcs in terms of net CO_2 inputs into the exogenic system (arc flux minus subduction flux). This is because much of the carbonate subducted may just come back out through the volcanoes, or as recently suggested, some of the carbonate subducted may bypass the volcanic front, given modern thermal states of subducting slabs (Dasgupta, 2013; Kerrick and Connolly, 2001). It is in continental arcs where the effect of upper plate carbonates may be most pronounced because continental margins serve as long-lived storage sites for sedimentary carbonate accumulated