Continental arc–island arc fluctuations, growth of crustal carbonates, and long-term climate change

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ABSTRACT

The Cretaceous to early Paleogene (ca. 140–50 Ma) was characterized by a greenhouse baseline climate, driven by elevated concentrations of atmospheric CO₂. Hypotheses for the elevated CO₂ concentrations invoke an increase in volcanic CO₂ production due to higher oceanic crust production rates, higher frequency of large igneous provinces, or increases in pelagic carbonate deposition, the last leading to enhanced carbonate subduction into the mantle source regions of arc volcanoes. However, these are not the only volcanic sources of CO₂ during this time interval. We show here that ocean-continent subduction zones, manifested as a global chain of continental arc volcanoes, were as much as 200% longer in the Cretaceous and early Paleogene than in the late Paleogene to present, when a cooler climate prevailed. In particular, many of these continental arcs, unlike island arcs, intersected ancient continental platform carbonates stored on the continental upper plate. We show that the greater length of Cretaceous–Paleogene continental arcs, specifically carbonate-intersecting arcs, could have increased global production of CO₂ by at least 3.7–5.5 times that of the present day. This magmatically driven crustal decarbonation flux of CO₂ through continental arcs exceeds that delivered by Cretaceous oceanic crust production, and was sufficient to drive Cretaceous–Paleogene greenhouse conditions. Thus, carbonate-intersecting continental arc volcanoes likely played an important role in driving greenhouse conditions in the Cretaceous–Paleogene. If so, the waning of North American and Eurasian continental arcs in the Late Cretaceous to early Paleogene, followed by a fundamental shift in western Pacific subduction zones ca. 52 Ma to an island arc–dominated regime, would have been manifested as a decline in global volcanic CO₂ production, prompting a return to an icehouse baseline in the Neogene. Our analysis leads us to speculate that long-term (>50 m.y.) greenhouse-icehouse oscillations may be linked to fluctuations between continental- and island arc–dominated states. These tectonic fluctuations may result from large-scale changes in the nature of subduction zones, changes we speculate may be tied to the assembly and dispersal of continents. Specifically, dispersal of continents may drive the leading edge of continents to override subduction zones, resulting in continental arc volcanism, whereas assembly of continents or closing of large ocean basins may be manifested as large-scale slab rollback, resulting in the development of intraoceanic volcanic arcs. We suggest that greenhouse-icehouse oscillations are a natural consequence of plate tectonics operating in the presence of continental masses, serving as a large capacitor of carbonates that can be episodically purged during global flare-ups in continental arcs. Importantly, if the global crustal carbonate reservoir has grown with time, as might be expected because platform carbonates on continents do not generally subduct, the greenhouse-driving potential of continental arcs would have been small during the Archean, but would have increased in the Neoproterozoic and Phanerozoic after a significant reservoir of crustal carbonates had formed in response to the evolution of life and the growth of continents.

INTRODUCTION

Earth’s surface temperature is controlled by the radiative energy balance of the atmosphere, which depends on insolation, albedo, and atmospheric composition, particularly the partial pressure of atmospheric CO₂ (pCO₂), a greenhouse gas (Kasting, 1993). Atmospheric pCO₂ is governed by exchange of C with other reservoirs of the exogenic C cycle (ocean and biosphere) and endogenic C cycle (crust and mantle) (Berner et al., 1983; Royer et al., 2004). On time scales >0.5 m.y., the exogenic system can be considered a single entity, with a mass of C dictated by inputs and outputs. Volcanism and metamorphism release CO₂, weathering produces HCO₃⁻, and deposition of organic matter and carbonate sequesters C (Berner, 1994; Kerrick, 2001; Ridgwell and Zeebe, 2005). The global cycling of C is completed on even longer time scales (>10 m.y.) when subduction of oceanic lithosphere transfers C into the deep Earth, and volcanic eruptions return this C to the atmosphere. Continental collisions also result in metamorphic decarbonation of limestones and in exposure and weathering of limestone and organic matter (Berner, 1994; Kerrick, 2001; Ridgwell and Zeebe, 2005). This is how the long-term C cycle is currently understood.

If long-term C sources and sinks were balanced, atmospheric pCO₂, averaged over millions of
of years, would remain relatively constant, and so would Earth’s surface temperature, unless insolation changes. This has not been the case over Earth’s history (Royer et al., 2004; Veizer et al., 2000) (Fig. 1A). Earth’s surface temperature changed significantly from the warm baseline of the Cretaceous and early Paleogene (ca. 130–50 Ma), which was characterized by relatively high sea level and a general lack of ice sheets (Hallam, 1984, 1985; Haq et al., 1987; Jenkyns et al., 2004; Müller and Groves, 1993; Forster et al., 2007; Wilson and Norris, 2001), to the cooler baseline of the late Paleogene to present (ca. 34–0 Ma), which was marked by large, though variable, ice sheets (Figs. 1A, 1B) (Hallam, 1985; Raymo, 1994; Zachos et al., 2008). Atmospheric \( p\text{CO}_2 \) is thought to have been 4–8 times higher during the Cretaceous and early Paleogene (1200–2400 ppm) than during the Holocene (~300 ppm) (Fig. 1B); this suggests that the warm baseline was driven by greenhouse conditions (Bice et al., 2006; Hong and Lee, 2012; Pagani et al., 2005; Pearson and Palmer, 2000; Royer et al., 2004). Based on these examples, the cycling of C between the endogenic and exogenic systems has clearly changed over long time intervals: the question is why.
CURRENT VIEWS ON AND QUESTIONS ABOUT CRETACEOUS–PALEogene GREENHOUSE

We use the Cretaceous–Paleogene greenhouse interval to explore possible causes of long-term climate variability. The most popular explanation for elevated atmospheric $p$CO$_2$ during this time interval centers on greater production of basaltic ocean crust through faster mid-oceanic ridge spreading rates and/or increased frequency of large igneous provinces (LIPs) (Fig. 1C; Berner and Kothavala, 2001; Hong and Lee, 2012; Kidder and Worsley, 2010; Larson, 1991a; Müller et al., 2008). This view appears consistent with the generally low $^{87}$Sr/$^{86}$Sr (Fig. 1D) and low Mg/Ca ratios of Cretaceous seawater (Fig. 1E), because faster spreading is thought to increase hydrothermal circulation rates through the ridge, which contributes unradiogenic Sr to seawater and consumes seawater Mg (Richter et al., 1992; Stanley and Hardie, 1998). Furthermore, at least two oceanic anoxic events (OAEs) during the Cretaceous have been attributed to eruption of LIPs and massive release of CO$_2$ (Adams et al., 2010; Kerr, 1998; Sinton and Duncan, 1997; Tarduno et al., 1991; Turgeon and Creaser, 2008).

However, it is unresolved whether oceanic crust production rates were sufficient to increase atmospheric $p$CO$_2$ by 4–8 times those of the present. To evaluate the contribution from oceanic crust production, we consider the following logic. The change in total C of the exogenic system (ocean + atmosphere) is given by

$$\frac{dC_{ex}}{dt} = j - k_{ex} C_{ex},$$  \hspace{1cm} (1)

where $dC_{ex}/dt$ is the rate (t = time) of change in concentration of C in the exogenic system, $j$ is the total CO$_2$ input into the exogenic system (volcanic degassing + weathering of organic C and carbonate), and $k_{ex}$ is the inverse response time for C sequestration from the exogenic system in the form of organic C and carbonate, and $C_{ex}$ is the concentration of C in the exogenic system. At the >1 m.y. time scales considered here, we can assume equilibrium between the ocean and the atmosphere. Then, following the assumption of Schrag et al. (2002) that the oceans are saturated in calcium carbonate, CaCO$_3$, on long time scales, equilibrium, $K$, between CO$_2$ in the atmosphere and ocean can be expressed as

$$K = \frac{[\text{Ca}^{2+}] [\text{HCO}_3^-]^2}{p\text{CO}_2},$$  \hspace{1cm} (2)

where [HCO$_3^-$] is the bicarbonate ion in the ocean, [Ca$^{2+}$] is the Ca$^{2+}$ concentration in the ocean and $p$CO$_2$ is atmospheric CO$_2$ concentration. Because the dominant C species in the ocean is HCO$_3^-$ (e.g., [HCO$_3^-$] $\sim$ $C_{ex}$) and the ocean-atmosphere system can be considered a steady state on time scales longer than 1/$k_{ex}$ $\sim$ 1 m.y. (Dessert et al., 2001), Equation 2 and the steady-state form of Equation 1 combine to yield atmospheric $p$CO$_2$ as a function of the input $j$ of CO$_2$ into the exogenic system:

$$p\text{CO}_2 \sim j^2 \left[ \frac{[\text{Ca}^{2+}]}{k_{ex} K} \right].$$  \hspace{1cm} (3)

If the quantity in parentheses is constant, $j$ scales approximately with the square root of atmospheric $p$CO$_2$. Therefore, increasing $p$CO$_2$ by 4–8 times requires a 2–2.8 times increase in CO$_2$ input into the exogenic system. In all likelihood, the rate constant $k_{ex}$ for CO$_2$ withdrawal from the system would increase during greenhouse conditions (because weathering efficiency would increase in a greenhouse world), so this 2–2.8 times increase in CO$_2$ output is a minimum.

What rates of oceanic crust production are needed to support such an increase in CO$_2$ output? Outgassing from ridges scales linearly with spreading rate if mantle potential temperature and average mantle C content are not changing substantially over the time scales of interest, as seems likely (Marty and Tolstikhin, 1998; McGovern and Schubert, 1989). This is because melting at ridges is driven by passive decompression, and consequently CO$_2$ flux through ridges scales as $FV_{Ca}$, where $C_{Ca}$ is the concentration of Ca in the mantle, $V$ is the half spreading rate, and $F$ is the average melting degree ($F$ is the flux of oceanic crust production), $C_{Ca}$ can be considered constant over 100 m.y. time scales. Perhaps more surprising is the observable fact that $F$, as inferred from oceanic crust thickness and geochemical tracers, is independent of spreading rate $V$ for all ridge segments except ultraslow spreading centers (Chen, 1992). However, the reasons for this are well understood on theoretical grounds: although the total amount of mantle passing through the melting regime beneath ridges will increase with increasing spreading, the fraction of such mantle that melts to form crust does not change (Langmuir et al., 1992). The only factor that can change $F$ is a change in mantle potential temperature, but this effect is not large enough. For example, increasing present-day mantle potential temperature of $\sim$1350 °C (Courtier et al., 2007; Lee et al., 2009; Putirka, 2005) by 100 °C would increase $F$ by at most -0.3 (see results of Katz et al., 2003). Lee et al. (2005) showed that mantle potential temperature beneath ridges may have been higher in the Cretaceous, but by no more than 100 °C (see fig. 9 in Lee et al., 2005). For these reasons, we expect magmatic flux of CO$_2$ through ridges to scale close to linearly with spreading rate. Thus, a 2–2.8 times increase in CO$_2$ input into the exogenic system, if driven solely by mid-oceanic ridge magmatism, requires an equivalent increase in global oceanic crust production rates.

Were oceanic crust production rates 2–2.8 times higher during the Cretaceous–Paleogene? Based on the areal distribution of seafloor ages, Rowley (2002) argued that global oceanic crust production rates were constant to within 20% over the past 200 m.y. In contrast, analyses of the same type of data by Larson (1991b) and Müller et al. (2008) suggest that global oceanic crust production rates may have been 1.5–1.75 times higher 50–140 m.y. ago compared to 0–50 m.y. ago. Demicco (2004) and Becker et al. (2009) suggested that 1.3–2 times higher rates in the Cretaceous are permitted by the seafloor-age distribution. All of these estimates of oceanic crust production are highly model dependent, so there is currently no consensus on how variable seafloor spreading was during the past 200 m.y. The highest estimated Cretaceous oceanic production rates may be sufficient to support Cretaceous greenhouse conditions, but the lower estimates are clearly insufficient. Until Cretaceous seafloor production rates can be better refined, additional sources of CO$_2$ should be explored.

A second possibility is to include the contribution from LIPs. However, given the brief durations of LIPs (<2 m.y.) and the <1 m.y. response time of CO$_2$ withdrawal from the exogenic system (Dessert et al., 2001), a LIP would have to erupt every ~1 m.y. to sustain prolonged greenhouse conditions for the time interval of interest. While LIP frequency was certainly higher in the Cretaceous–Paleogene than the mid-Cenozoic, the frequencies are still far too low to sustain continuous and prolonged greenhouse conditions (Fig. 1C) (Müller et al., 2008).

A third scenario invokes an increase in pelagic carbonate deposition during the Cretaceous, possibly facilitated by the evolutionary rise of pelagic carbonate-secreting organisms in the early Mesozoic (Katz et al., 2004). Such a scenario would increase the delivery of carbonate to subduction zones, from which it follows that the output of CO$_2$ through arc volcanoes would increase (Caldeira and Rampino, 1993; Edmond and Huh, 2003; Johnston et al., 2011). However, it is not obvious that enhanced sequestration of carbonate in oceanic crust would necessarily increase the total flux of CO$_2$ out of arc volcanoes. Consider a modified form of Equation 1, which describes the change in C content of the exogenic system $C_{ex}$.
\[
\frac{dC_{ex}}{dt} = j - k_{oc,ex}C_{oc} + \left( k_{ex,cc}C_{ex} - k_{ex,oc}C_{ex} \right),
\]
(4)

where \( \frac{dC_{ex}}{dt} \) is the rate of change of C in the exogenic system, \( C_{ex} \) is the concentration of C in the subducted oceanic crust, \( k_{ex,cc} \) is the rate constant for sequestration of exogenic C into non-subductable C on continental platforms, \( k_{oc,ex} \) is the rate constant describing the transfer of metamorphically decarbonated C from the subducted slab back to the exogenic system through arc volcanism, and \( k_{ex,oc} \) is the rate constant describing sequestration of exogenic C into oceanic crust in the form of carbonated sediments or veins in the oceanic crust. If we consider the case in which 100% of subducted carbonate is completely decarbonated and returned to the exogenic system through arc volcanoes, then the quantity in parentheses on the right side of Equation 4 equals zero, regardless of changes in the efficiency of pelagic carbonate deposition, e.g., if \( k_{oc,ex} \) varies. With 100% decarbonation of the subducted oceanic lithosphere, the steady-state concentration of C in the exogenic system, \( C_{ex} \), must then be given by \( j/k_{ex,cc} \). It is clear that \( C_{ex} \) must ultimately be driven by external forcings, i.e., \( j \). The subduction factory, in the context of the foregoing hypothesis, is not an external forcing. Finally, metamorphic decarbonation in typical subducting slabs is likely to be far less than 100% efficient because the temperatures required for decarbonation increase with increasing pressure (Dasgupta and Hirschmann, 2010; Kerck and Connolly, 2001; Tsuno and Dasgupta, 2012). For example, in Figure 2, it can be seen that decarbonation, even in the presence of silica and H\(_2\)O, is unlikely to be extensive because typical slab temperatures are not hot enough. Thus, enhanced pelagic deposition of carbonates, in the absence of an external forcing of CO\(_2\), could even lead to a net decrease in exogenic C.

We can conclude that an increase in external forcing is necessary to drive elevated atmospheric pCO\(_2\) in the Cretaceous–Paleogene. While juvenile mantle inputs seem to be favored by the community, they may not, by themselves, be sufficient. One remaining possibility is to liberate CO\(_2\) from long-lived reservoirs of car-

Figure 2. (A) Subsolidus decarbonation reactions between dolomite (Do), calcite (Cc), and quartz (Qz) to form diopside (Di), wollastonite (Wo) and CO\(_2\). \( P \)—pressure; \( T \)—temperature. Dark dashed curve is a steady-state conductive geotherm for an 80 mW/m\(^2\) surface heat flow and a 35-km-thick crust. Dark solid line is a transient geotherm generated during continent-continent collision (Huerta et al., 1999). Slab surface temperatures (patterned field) in global subduction zones are from Syracuse et al. (2010). Gray shaded vertical bar is the range of magmatic temperatures. (B) Decarbonation reactions at 0.3 (lower) and 1 GPa (upper) as a function of CO\(_2\) activity. Temperature of decarbonation decreases with decreasing aCO\(_2\), such as might occur under H\(_2\)O-rich conditions. Gray horizontal bar corresponds to possible magmatic temperatures.
continental arcs and climate

Based on a compilation of geologic maps (see Appendix), continental arcs (e.g., Andean style volcanism) were longer in the Cretaceous than in the mid-Cenozoic (Figs. 3A and 4). Cretaceous–Paleogene continental arcs were longer in the Cretaceous–early Paleogene than in the mid-Cenozoic (see Appendix), continental arcs (e.g., Andean island arcs (e.g., the modern Marianas and outer Aleutians) are built on oceanic lithosphere. Because carbonate production and preservation are favored in shallow waters, most of the global marine carbonate accumulation, at least until the Mesozoic, occurred over the relatively small area of continental shelves and slopes above continental or transitional crust rather than across the enormous expanse of the deep sea above oceanic crust (Wilkinson and Walker, 1989). Thus, continental arc magmas are more likely than island arc magmas to intersect sedimentary carbonates (Fig. 5). Because of the high temperatures (600–1200 °C) characteristic of magmas (Fig. 2), intrusion of magmas into carbonates can result in efficient decarbonation by contact metamorphism, assimilation, or metamorphic reaction with the magma. During decarbonation, every mole of carbonate (limestone or dolomite) that reacts with magma releases 1 mol of CO2 (CaCO3 + SiO2 = CaSiO3 + CO2).

CONTINENTAL ARCS DOMINATED IN THE CRETACEOUS

There are several implications. Continental arcs are built through continental crust, whereas island arcs (e.g., the modern Marianas and outer Aleutians) are built on oceanic lithosphere. Because carbonate production and preservation are favored in shallow waters, most of the global marine carbonate accumulation, at least until the Mesozoic, occurred over the relatively small area of continental shelves and slopes above continental or transitional crust rather than across the enormous expanse of the deep sea above oceanic crust (Wilkinson and Walker, 1989). Thus, continental arc magmas are more likely than island arc magmas to intersect sedimentary carbonates (Fig. 5).

Figure 3. (A) Paleogeography during the Late Cretaceous (Ulmishek and Klemme, 1990). Red triangles are known continental arcs between 140 and 60 Ma (see Appendix for references). Blue represents flooded regions of continents (Ulmishek and Klemme, 1990). Black regions are Cretaceous hydrocarbon source rocks (Ulmishek and Klemme, 1990). (B) Distribution of Phanerozoic carbonates older than the Cretaceous. Precambrian carbonates are not shown, so this is a lower bound on the distribution of crustal carbonates.
requires interaction with carbonate wall rock in crustal magma chambers, indicating that at least some component of carbonate is derived from the upper plate.

To place carbonate-intersecting volcanoes like Etna and Vesuvius in context, we examine their contributions to the global CO₂ output through volcanoes. Etna (crater summit and flank diffusion) puts out ~6 × 10¹¹ mol/yr based on SO₂/CO₂ measurements of degassed volatiles and SO₂ fluxes (Allard et al., 1991). For comparison, the present-day global CO₂ emission through subduction zones is ~1.5–3.1 × 10¹² mol/yr (Dasgupta and Hirschmann, 2010; Hilton et al., 2002; Marty and Tolstikhin, 1998; Sano and Williams, 1996), a flux primarily based on circum-Pacific volcanoes, which are mostly island arcs or juvenile continental arcs and where the amount of crustal carbonate is comparatively smaller. The present-day CO₂ emission through mid-oceanic ridges is ~1–5 × 10¹² mol/yr (Dasgupta and Hirschmann, 2010; Marty and Tolstikhin, 1998; Sano and Williams, 1996). Subduction zones (excluding volcanoes, such as Etna and Vesuvius) and ridges collectively yield a CO₂ output of ~4–10 × 10¹² mol/yr (Marty and Tolstikhin, 1998). Thus, Etna alone accounts for 10%–14% of the global CO₂ output. Those volcanoes that produce the most CO₂ thus appear to intersect thick sections of crustal carbonates, suggestive of a causal link between high CO₂ production and magmatic emplacement into carbonate platforms. Examples of other high CO₂-producing volcanoes, though less productive than Etna, include Merapi in Indonesia, where skarn xenoliths and zoned phenocrysts indicate magma interaction with carbonate wall rock in shallow magma chambers (Chadwick et al., 2007; Deegan et al., 2010; Toutain et al., 2009). Although more work is necessary to evaluate the relative contributions of crustal carbonate...
versus subducted carbonate to the flux of CO₂ emitted by arc volcanoes, the examples indicate that crustal assimilation in magma chambers may play an important role when carbonates exist in the upper plate. We note that only 50 or so Etna-type volcanoes would increase the global CO₂ volcanic output by a factor 3–7 relative to the present-day global output, more than enough to drive the Cretaceous greenhouse. In the following we consider a scenario in which the Cretaceous–early Paleogene was characterized by a global chain of carbonate-intersecting arc volcanoes.

**EVIDENCE FOR WIDESPREAD MAGMATIC DECARBONATION OF UPPER PLATE CRUSTAL CARBONATES**

We attempt to estimate the extent to which Cretaceous to early Paleogene continental arc volcanoes may have intersected crustal carbonates in the upper plate. For a conservative assessment of this possibility, we mapped the distribution of Phanerozoic carbonates deposited before the Cretaceous (Fig. 3B). Where batholiths were emplaced in Tibet, southeastern Asia, eastern China, and Korea, thick Paleozoic sedimentary sections containing carbonates covered much of the prebatholithic crust (Golonka et al., 2006). Mesozoic batholiths in the North American Cordillera appear to have intersected Proterozoic and Paleozoic sediments, as evidenced by abundant roof pendants in the region (Burnham, 1959; Kerrick, 1970; Lackey and Valley, 2004; Pickett and Saleeby, 1993). In some of these regions, sedimentary sections approach 10 km in thickness, and limestones and dolostones occur in >50% of the stratigraphy (Stevens and Greene, 1999). Few Paleozoic carbonates crop out in western South America, so the contribution of the Andean volcanism to decarbonation may have been small both in the past and the present. We estimate the length of continental arcs that may have intersected crustal carbonates to be ~13,000 ± 2000 km, ~25% of the total length of modern arcs. We use this estimate in the following discussion to predict how much CO₂ may have been liberated.

We can further evaluate the areal extent of skarn formation by mapping the distribution of decarbonated residues, i.e., skarns. Skarns are seldom represented in large-scale geologic maps because of their heterogeneous nature; however, because they are often associated with strategic metal ores (W, Sn, Cu, and Zn) (Barton and Hanson, 1989), their distribution can be determined with some confidence. For example, scheelite (CaWO₄) forms exclusively in carbonate skarns when the tungstate ion (WO₄²⁻) in the silicate magma replaces the carbonate ion (CO₃²⁻) in carbonate, liberating CO₂ (Wan, 1987). The use of W alloys in armor-piercing warheads made W a strategic metal, particularly during World War II, so locations of scheelite deposits across North America are established (Albers, 1981; John and Bliss, 1993; Newberry and Einaudi, 1981; Sinclair, 1995; Werner et al., 1998). Not all skarns contain scheelite, and many skarns have likely been eroded, hence the distribution of scheelite is a minimum map extent of skarn formation. In Figures 6 and 7 we show the distribution of Cretaceous–Paleogene plutonic rocks in North America and the distribution of Cretaceous–early Paleogene skarns. The abundance of Cretaceous–Paleogene scheelite deposits throughout the length of contemporaneous batholiths in North America qualitatively suggests that extensive decarbonation of crustal carbonates may have taken place in these Cretaceous–early Paleogene continental arcs.

We can also shed insight on the vertical extent of skarn formation. Skarn formation in the Sierra Nevada has been documented at near-surface conditions and in the deep crust.
Figure 7. Distribution of scheelite deposits (and some Sn-W and Cu-Mo skarns) of Cretaceous–early Paleogene age. Data sources are provided in the Appendix.

For example, oxygen isotopic studies of Sierran skarn outcrops suggest pervasive infiltration of surface-derived waters, which implies formation conditions at <1 km depth, providing ideal conditions for decarbonation (D’Errico et al., 2012). Evidence for lower crustal decarbonation comes from scheelite-bearing skarns and calc-silicate metamorphic rocks in the southern Sierra Nevada, where intrusive rocks derived from paleodeeps of 20 km have been exhumed (Ague and Brimhall, 1988; Pickett and Saleeby, 1993) (Fig. 7). These deep-seated skarns and/or endoskarns likely derive from the underthrusting of craton-derived sediments into the magmatic arc during Cretaceous time (DeCelles et al., 2009) (Fig. 5B). Magmatic assimilation of carbonated metasediments would be a likely source of mid-crustal decarbonation. In the Peninsular Ranges batholith (California), the protoliths of many enclaves in the plutons are crustal carbonates or carbonate-bearing sandstones, now manifested in a spectrum of lithologies including wollastonite-, diopside-, amphibole- or biotite-rich rocks (Fig. 8). In addition, pluton margins in contact with carbonate country rock are often metasomatized into endoskarns, often manifested in the form of plagioclase-quartz-pyroxene rocks representing the product of Ca and Mg metasomatism of the pluton bodies (Dyer et al., 2011) (Fig. 8).

Skarn formation associated with continental arc magmatism was thus pervasive throughout the North American Cordillera in both lateral and vertical extent during the Cretaceous and early Paleogene. The North American Cordillera represents only one-third of the Cretaceous–early Paleogene continental arcs that intersect crustal carbonates (see Figs. 3, 6, and 7). Existing global data sets suggest that this pattern of skarn formation also applies to Eurasia; some of the world’s largest scheelite deposits and associated skarns occur in Korea and southeast China, where late Mesozoic granitoids have intruded carbonates (Werner et al., 1998). Today, few arcs intersect carbonates, and as mentioned earlier, prebatholithic carbonates are few in the Andean arc. The likelihood of profound release of CO₂ from magmatically induced decarbonation of crustal carbonates during the Cretaceous–early Paleogene was clearly high. We attempt to quantify this contribution of CO₂ in the next section.

We emphasize that continental arc environments are more efficient at decarbonating crustal carbonates than continent-continent collisions (Kerrick and Caldeira, 1994) or decarbonating subducting lithosphere. This is because, in continental arcs, (1) magmatic intrusions advect heat into the crust, which elevates the geotherm (Fig. 2A), (2) temperatures of decarbonation are low at upper crustal pressures, and (3) high water/rock ratio hydrothermal circulation is more pronounced in the upper crust because of the abundance of meteoric waters as well as higher permeability. Decarbonation must be less efficient in continental collisions and in subducting slabs because the magmatic flux is small to negligible, the geotherm is depressed, and decarbonation temperatures are higher (due to higher pressures involved; Fig. 2A) (Huerta et al., 1999).

IMMEDIATE CONSEQUENCES OF A CONTINENTAL ARC–DRIVEN CLIMATE

There are several testable consequences of a greater length of continental arcs and a more expansive area of magma-carbonate interaction.

Enhanced CO₂ Fluxes

The best way to estimate CO₂ production from magmatically induced decarbonation of wall rock and assimilated sediments is to quantify the volume of carbonated sediments that have interacted with plutons and the extent to which decarbonation reactions have progressed.
Continental arcs and climate

Although we showed that Cretaceous–Paleogene skarn deposits are ubiquitous, we are far from quantifying volumes and extents of decarbonation for each skarn deposit so that such data can be assimilated on a global scale. Nevertheless, we can obtain a crude estimate of the relative increase in CO2 production when carbonate-intersecting continental arcs are active. We do this by describing the volcanic input of CO2 (expressed as the mass of CO2/time), j, as a perturbation to volcanic CO2 production in the late Cenozoic:

\[ j = J \left( C_o + LX C_o \right), \]  

(5)

where \( J \) is the global magmatic production rate of arc magmas, \( C_o \) is the average concentration of CO2 in primitive undegassed arc magmas today, \( L \) is the additional fractional length of carbonate-intersecting continental arcs in the Cretaceous–Paleogene relative to the present, \( X \) is the mass fraction, relative to the mass of magma, of pure carbonate sediment assimilated into magmas or decarbonated in skarns, and \( C_{co} \) is the concentration of CO2 in limestones and dolostones (44–48 wt%). The quantity \( J C_o \) represents the present-day baseline of global CO2 output from arc volcanoes \( j_o \), Values for \( C_o \) are highly uncertain because CO2 degasses readily at depth. An upper bound of 0.3–0.5 wt% for primary undegassed concentrations is indirectly inferred from the global CO2 fluxes from arc volcanoes, but estimates from melt inclusions are lower by an order of magnitude (Wallace, 2005). If we assume that global magmatic production of arc magmas remains constant and that the only change is in the proportion of island and carbonate-intersecting continental arcs, we can reexpress Equation 5 as a quantity relative to present-day volcanic inputs:

\[ \frac{j}{j_o} = 1 + \frac{LX C_o}{C_o}, \]  

(6)

where \( \frac{j}{j_o} = J C_o \) and accounts for the present-day global arc production of CO2.

We can calculate a minimum amount of CO2 produced during crustal decarbonation by constraining the amount of sedimentary carbonate directly assimilated into the magma. Oxygen isotopic compositions of evolved granitoid plutons in continental arcs are as high as \( +10\%e \) relative to modern seawater, whereas juvenile mantle-derived magmas are \( +5\%e \) (Hill et al., 1986; Kistler et al., 2003; Mattey et al., 1994). Assuming that metasediments have oxygen isotopic compositions in excess of \( +20\%e \), it is not unreasonable for evolved granitoids to have assimilated >30% sediments or partial melts of sediments. Such estimates are consistent with those based on \( ^{87} \text{Sr} / ^{86} \text{Sr} \) (Kistler et al., 1965). If carbonates make up, on average, one-third of platform sedimentary packages (Ronov, 1972), >10% assimilation of pure carbonate would be implied. Similar carbonate fractions are seen in the Paleozoic strata on the western margin of North America, through which the Cordilleran plutons intruded (Greene and Stevens, 2002; Stewart, 1972). Taking the additional length of carbonate-intersecting continental arcs \( L \) in the Cretaceous–Paleogene as \( 25\% \) and maximum \( C_o \) as 0.3–0.5 wt%, Equation 6 yields 3.1–4.6 times more production of CO2 than the present. This is a lower bound, because assuming a lower value for \( C_o \) would increase the relative magnitude of Cretaceous–Paleogene CO2 production. In addition, skarn formation driven by hydrothermal fluid circulation in the contact aureole surrounding plutons would only add to these estimates (Kerrick and Caldeira, 1998; Nesbitt et al., 1995). In summary, if all of this CO2 escaped to the atmosphere, this would be more than sufficient to increase atmospheric \( p \text{CO}_2 \) by the 4–8 times. If some of this CO2 is trapped in the crust or plutons in the form of secondary calcite veins, the estimated CO2 production would be lower. Carbonates, interpreted to be primary, are commonly observed in granitoid plutons, but at the <1% level (White et al., 2005). The pervasive occurrence is consistent with widespread fluxing of CO2 through continental arc magmas, but the trace levels suggest insignificant retention.

**Enhanced Production of Marine Carbonate and Organic C**

Higher atmospheric \( p \text{CO}_2 \) in the Cretaceous to early Paleogene, if driven by enhanced CO2 production, should lead to enhanced outputs of
carbon from the exogenic system, provided that burial efficiency keeps up with CO₂ production. This is consistent with evidence for widespread deposition of carbonate platforms and extensive hydrocarbon source rocks during the Cretaceous and early Paleogene (Fig. 3A). Although higher pCO₂ should decrease ocean pH on short (<10,000 yr) timescales, elevated CO₂ emissions on long (>10,000 yr) timescales should enhance carbonate accumulation (Ridgwell and Zeebe, 2005; Leon-Rodriguez and Dickens, 2010). That is, if total carbon inputs to the exogenic carbon cycle are large, then the outputs from the exogenic system in the form of carbonates and organic C should be similarly large. Inputs and outputs balance on long time scales. This may explain the large quantities of limestones and black shales (and hydrocarbon source rocks) deposited during the Cretaceous and early Paleogene. It also follows that the carbonate compensation depth, i.e., the depth at which the sinking rate of carbonate equals the dissolution rate, may have deepened in the Cretaceous to early Paleogene, allowing more widespread deposition and preservation of carbonate in pelagic environments compared to the mid-Cenozoic to present. Pelagic carbonate deposition would be a consequence of higher inputs of CO₂ into the exogenic system rather than the driver of higher inputs into the exogenic system.

**Carbon Isotopic Composition of Volcanic CO₂ Is Increased**

Limestone and dolostone decarbonation would have enriched total volcanic inputs in δ¹³C compared to typical mantle-derived CO₂. This would cause an overall positive shift in the δ¹³C of the exogenic C cycle, consistent with the higher δ¹³C values for Cretaceous–early Paleogene marine carbonate relative to the late Cenozoic (Fig. 1D). For example, a 2%–3% increase in seawater δ¹³C can be generated by a crustal carbonate decarbonation flux of CO₂ equivalent to 50%–100% of the modern background emissions of CO₂ into the atmosphere (Fig. 9; see Appendix for calculation details). For comparison, the same increase in seawater δ¹³C can be generated by increasing the proportion of organic C burial to 0.3–0.35 from 0.2 in the late Cenozoic or by decreasing the isotopic fractionation factor of C during organic C burial relative to seawater by 1%–1.5%. This calculation is more complicated if organic C, in addition to carbonate, is liberated as CO₂ to the atmosphere during magma-sediment interaction. What is clear is that the C isotopic composition of inputs into the exogenic system is unlikely to be equal to canonical mantle.

Enhanced Ash Production

Continental arcs are also more explosive than island arc, ridge, and LIP volcanism due to their more evolved (andesite) and volatile-rich compositions. Continental arc volcanoes can thus spew fine ash into the high atmosphere, distributing it globally on time scales ranging from days to a few years (McCormick et al., 1995). On longer time scales (10–100 k.y.), physical and chemical weathering of the volcanic slopes would prolong the transport of particulates and solutes of volcanic origin into the ocean. This prediction is consistent with widespread Cretaceous bentonite deposits (altered ash) throughout western North America, as far as the modern Mississippi River, more than 2000 km east of the Cordilleran arc (Dawson, 2000; Elder, 1988; Nixon, 1973). It is interesting that during modern eruptions, marine phytoplankton have been shown to assimilate Fe and other metals from ash within a few days of eruption (Duggen et al., 2007), begging the question of whether organic C and ash deposition may have been closely associated in the Cretaceous to early Paleogene. It is possible that widespread ash deposition aided the diversification of diatoms and radiolarians, resulting in organic-rich chert deposits in some parts of the Pacific Ocean (Dumitrescu and Brassell, 2005) (Fig. 1F).

³⁶Sr/⁸⁶Sr of Seawater

We discuss here the possible impacts of a continental arc–derived world on seawater tracers, such as ³⁶Sr/⁸⁶Sr and Mg/Ca (Fig. 1). The ³⁶Sr/⁸⁶Sr of continental arc magmas should be intermediate between unradiogenic and radiogenic Sr, the former driven by juvenile magmatism and the latter controlled by assimilation of ancient continental crust components. Weathering of continental arcs might be expected to slightly increase the ³⁶Sr/⁸⁶Sr of seawater; this is seemingly inconsistent with the notably unradiogenic signature (0.7072) of Cretaceous–Paleogene seawater, and would seem to require enhanced oceanic crust production to compensate for this increase. However, the main orogenic belts during the Cretaceous–Paleogene were associated with continental arcs, whereas the orogenies in the mid-Cenozoic were characterized by continent-continent collisions (e.g., closing of the Tethys Ocean). Sr derived from weathering of collisional orogenies should be more radiogenic than that derived from continental arcs, and hence the Sr isotopic composition of the global continental weathering flux should have been lower in the Cretaceous–Paleogene seawater, and would seem to require enhanced oceanic crust production to compensate for this increase. We can quantify this effect by considering the following. Assuming steady state, seawater Sr isotopic composition is a mixture of continental-derived weathering flux and oceanic crust-derived hydrothermal flux:

\[
R_w = R_{cw} \left(1 - X_w\right) + R_{oc} X_w, \tag{7}
\]

where \(R_w\), \(R_{cw}\), and \(R_{oc}\) are the ³⁶Sr/⁸⁶Sr of seawater, continental crust, and oceanic crust, and \(X_w\) is the mass fraction of Sr in the ocean.
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Mg/Ca of Seawater

Seawater Mg/Ca has traditionally been interpreted to reflect enhanced hydrothermal circulation of seawater through oceanic crust, resulting in the sequestering of Mg (Stanley and Hardie, 1998). One might be tempted to suggest that skarn formation could drive a fundamental change in Mg/Ca ratio of river waters and hence seawater. However, any carbonate-derived Ca and Mg would become locked in silicates during skarnification and would therefore not be simultaneously liberated with CO₂ to the exogenic system. Subsequent weathering of skarns, if dominated by Ca skarns, could increase Ca input into the oceans, thereby decreasing seawater Mg/Ca. However, this effect would most certainly be diluted to some extent by the background weathering of igneous rocks and noncarbonate sedimentary rocks. We therefore do not expect to see any significant effects of enhanced skarn formation on seawater Mg/Ca.

However, it is important to note that continental arcs are often associated with dynamic subsidence of the continental plate, resulting in the development of extensive shallow seaways (Gurnis, 1992; L. Liu et al., 2008; S. Liu et al., 2011) (Fig. 3). Burns et al. (2000) suggested that times of low seawater Mg/Ca seem to have coincided with times of enhanced dolomite formation in shallow continental interior seaways. This begs the question of whether low seawater Mg/Ca in the Cretaceous–early Paleogene could have in part been driven by enhanced sequestration of Mg during diagenetic dolomite formation (Wilkinson and Algeo, 1989). At the very least, such a scenario should be considered in addition to the conventional view of hydrothermal sequestration in oceanic crust. Mg and Ca isotopic proxies of Cretaceous seawater might provide a tool for testing between these ideas.

FLUCTUATIONS BETWEEN CONTINENTAL AND ISLAND ARC STATES

A fundamental question is why continental arc magmatism was more prevalent during the Cretaceous and early Paleogene. This might be understood by considering different dynamics involved in ocean-ocean and ocean-continent subduction zones, the former generating island arcs and the latter generating continental arcs (Stern, 2002; Uyeda and Kanamori, 1979) (Fig. 10). In subduction zones where the upper plate responds passively to the rollback of the subducting slab, the upper plate extends, resulting in backarc rifting, the formation of a marginal basin, and the development of an island arc. If, however, the upper plate is actively pushed toward the trench or if slab rollback is impeded, the upper plate is placed into compression, closing the marginal basin to form a continental arc. Unlike backarc spreading behind intraoceanic subduction zones, the backarc regions in ocean-continent subduction zones are under compression, leading to large-scale folding, thrusting, and the development of large basins driven by dynamic subsidence (rather than by lithospheric thinning). This compression regime is exemplified in the modern Andes and in the Cretaceous–early Paleogene North American Cordillera (DeCelles, 2004; DeCelles et al., 2009).

It is possible that global changes in the nature of subduction zones are consequences of the aggregation and dispersal of continents (Fig. 10). When continents are aggregated (e.g., Pangaea), vertical heat loss is decreased through the stagnant continental lid and lateral heat loss is decreased by the presence of cold subducting oceanic lithosphere surrounding the continents (Gurnis, 1988; Lenardic et al., 2011). In con-

Figure 10. Speculative conceptual model for fluctuations between continental and island arc states. (A) Backarcs of subduction zones are in extension, favoring island arcs. CO₂ emitted during this time is derived from the mantle and from subducted carbonate. An island arc scenario might be favored during continent aggregation. During this time, the continental mantle dome heats and the oceanic domain cools, setting the stage for a large mantle overturn event. (B) Mantle overturn triggers continental dispersal, driving the leading edge of continents to advance trenchward, which places subduction zones into compression. Continental arcs and foreland fold and thrust belts in the backarc region ensue. During the continental arc stage, previously stored carbonate is purged as CO₂, amplifying total volcanic emissions of CO₂ as well as total deposition of carbonate and organic C.
and purging of CO₂ (continental arc stage) (Fig. 2). Between periods of C storage (island arc stage) on time scales of continent assembly and dispersal, the Earth’s history should have fluctuated, on the ~200 m.y. timescale. Icehouse baselines define the former and greenhouse states may be characteristics of Earth’s climate. There are reasons to suspect that the size of the crustal carbonate reservoir has not been constant through time. Carbonates, being preferentially deposited on continental shelves and slopes, have a low probability of being subducted. This means that the mass of carbonates in the continents has likely grown with time, dictated in part by the evolution of life and the growth of the continents upon which the carbonates are deposited and stored (Ronov, 1972). It follows that long-lived tectonically driven fluctuations between greenhouse and icehouse states may be characteristics of Earth’s recent past (particularly after the Cambrian explosion of life) and perhaps absent early in Earth’s history.

**IMPLICATIONS FOR LONG-TERM CLIMATE VARIABILITY**

Fluctuations between island arc– and continental arc–dominated states, if linked to the aggregation and dispersal of continents, must have operated for as long as plate tectonics have operated and continents have existed. Provided that significant crustal carbonate reservoirs have always existed, the CO₂ content of the atmosphere should have fluctuated, on the ~200 m.y. timescale of continental assembly and dispersal, between periods of C storage (island arc stage) and purging of CO₂ (continental arc stage) (Fig. 10); icehouse baselines define the former and greenhouse baselines define the latter.

If the Cretaceous–early Paleogene greenhouse represents a continental arc–driven world, the icehouse conditions of the mid-Cenozoic to present result from a return of the

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**APPENDIX. DISCUSSION OF METHODS AND RESULTS**

**Methods**

Cretaceous–Paleogene paleogeography with regions of continental inundation and hydrocarbon source rocks in Figure 3 is adapted from Ulmishek and Klemme (1990). Continental arcs from this time were compiled from a number of references. Cretaceous–Paleogene batholiths occur along the (1) entire stretch of the North American Cordillera, extending from southern Alaska south through western North America and into Mexico (Barton, 1996; Coleman and Glazner, 1997; Kiesler et al., 2003), (2) the South American Cordillera, extending from Columbia south to southern Chile (Kay et al., 2004), (3) the southern margin of Eurasia extending from eastern Turkey east through southern Tibet and into present-day southeastern Asia (Bozüyak et al., 2007; Golonka et al., 2006; Wen et al., 2008), and (4) the eastern part of Eurasia, extending from southern China (Li, 2000), north through Korea, Japan, and eastern Siberia (Akinin et al., 2009; Chough and Sohn, 2010; Kagami et al., 1992; Miller et al., 2002; Nakajima, 1996). The total length of continental arcs was estimated using the ruler tool in Google Earth. We also present the locations of Phanerozoic carbonates prior to the Cretaceous using the global carbonate map and paleogeographic study of Kissling et al. (2003). This work was supplemented by that of Sengör and Natal’in (1996) and Meyerhoff et al. (1991) for China and Tibet, Golonka et al. (2006) for southeast Asia, Jaillard et al. (2000) and Ramos (2000) for South America, and for North America and Central America by Cook and Bally (1975). The skarn database was based on the following sources. We first used literature compilations of scheelite, Sn-Mo, and Cu-Au skarns (Albers, 1981; John and Bliss, 1993; Newberry and Einaudi, 1981; Sinclair, 1995; Wernert et al., 1998). These were amplified by extracting data on unpublished scheelite localities in the western United States using the U.S. Geological Survey Mineral Resources Data System (http://tn.er.usgs.gov/mrds/) and in British Columbia from the British Columbia Ministry of Energy mineral resources database MINFILE (http://minfile.gov.bc.ca/searchbasic.aspx). Additional skarn deposits from Alaska were taken from Newberry et al. (1997). Only Cretaceous and Paleogene skarns were selected for presenting in the text. Skarn ages correspond to the age of the magmatic intrusion, not to the age of the protolith or country rock. In some cases, pluton ages are described in the published literature. In many other cases, pluton ages are only shown in geologic maps. We correlated each skarn locality to the host pluton using geologic maps using OneGeology (www.onegeology.org), an online database of geologic maps based on the Geological Society of America Decade of North American Geology base map (Reed et al., 2005). The filtered skarn data were combined with the Geological Society of America base map in ArcGIS (geographic information system; http://www.esri.com) to produce Figure 7.

Strontium, carbon, and sulfur isotope data in Figure 1 are from the published database of Prokop et al. (2008). Because we are interested in long-term (>10 m.y.) variations in these isotopic signatures of seawater, a running average based on 10 m.y. intervals was calculated.

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Mass Balance Estimates

To estimate how magmatic decarbonation of crustal carbonates affects the C isotopic composition of the exogenic system and seafloor, a simple mass balance is considered here. Under steady-state conditions, inputs and outputs must be balanced:

\[ R_{xw}F_{xw} + R_{vw}F_{vw} = R_{sw}F_{sw} + R_{oo}F_{oo} \]  

(A1)

where \( R_{xw} \) and \( F_{xw} \) are the relative proportions of the CO2 inputs to the exogenic system from magmatically induced decarbonation of crustal carbonates (e.g., limestones and dolostones) in continental arcs and the background flux of CO2 from ridges, island arcs, and metamorphism and weathering of organic C and carbonates on continents, respectively. \( R_{vw} \) and \( F_{vw} \) are the relative proportions of CO2 outputs in the form of organic C burial and carbonate burial, the latter associated with the end product of chemical weathering of silicates, \( R_{sw} \) and \( F_{sw} \) are the isotopic ratios of \(^{13}C/^{12}C\) in crustal carbonates, mantle-derived CO2, organic C, and seawater, the latter of which we take, for simplicity, to be equivalent to carbonates precipitated from seawater. The isotopic ratio is related to \(^{13}C/^{12}C\) notation as \( R = R_{ref}^{13C} + \Delta R \), where \( R_{ref} \) represents the ratio of a standard, typically PDB or SMOW.

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(a similar equation can be written for \( F_{oo} \)). Equations A4 and A5 can be modified algebraically to yield

\[ R_{sw} = \frac{R_{vw} + j_{vw} / j_{sw} \cdot R_{oo} \cdot R_{vls}}{(j_{sw} / j_{oo}) R_{oo} + R_{sw}} \cdot \Omega \]  

(A7)

Equation A7 shows how much magmatic decarbonation of crustal carbonates is needed to change the C isotopic composition of seawater \( R_{sw} \) from a reference state \( R_{sw}^{ref} \). Note that \( R_{sw} \) is assumed constant and equal to the canonical C isotopic composition of the mantle given in delta notation as \( \Delta R_{C} = -5 \)‰ (this will be the same if the flux of CO2 from weathering of organic C and carbonate in continents is not \(-5 \)‰), \( R_{oo} \) is the average C isotopic composition of crustal carbonates (0‰ to +1‰), and \( j_{vw} \) is assumed to be a baseline flux of CO2 that does not change over the time interval of interest (because we are only assessing the influence of magmatic decarbonation of crustal carbonates). The latter assumption implies then that \( j_{vw} / j_{sw} \) represents the excess CO2 inputs into the atmosphere associated with magmatic decarbonation of crustal carbonates in continental arcs. If \( j_{vw} / j_{sw} = 0 \), i.e., there are no contributions from crustal carbonate decarbonation, then Equation A7 simplifies to

\[ R_{sw} = \frac{R_{vw}}{R_{vls}} \cdot \Omega + R_{oo} \cdot \frac{R_{oo} - R_{sw}}{R_{oo} - 1} \]  

(A8)

Thus, if \( F_{sw} \) and \( \alpha \) remain constant and \( j_{sw} \) is zero, \( R_{sw} \) does not vary with time, i.e., \( R_{sw} / R_{sw}^{ref} = 1 \).

The sensitivity of \( R_{sw} \) to changes in \( F_{sw} \), \( \alpha \), and \( j_{vw} / j_{sw} \) is assessed as follows. We take the present as a reference, so we assume that \( j_{vw} / j_{sw} \) is close to zero today. In delta notation, the C isotopic composition of seawater, \( R_{sw} \approx 0.975 \), is equivalent to 0‰ (equivalent to a \(^{12}C/^{12}C\) of 0.011237), \( R_{oo} \approx -5 \)‰, and the isotopic composition of modern organic C is \(-25 \)‰. This gives a modern fractionation factor between organic C and seawater of \( \alpha \approx 0.975 \). Assuming that carbonate precipitation does not fractionate C isotopes, these values correspond to a modern organic C burial proportion of \( F_{sw} = 0.2 \), or 20% of the total deposition of C today is in the form of organic C and the remaining is in the form of carbonate. This is consistent with the analysis by Shackleton (1987).

Cretaceous–Paleogene seawater was characterized by C isotopic compositions that were 2‰–3‰ higher than the late Cenozoic. As can be seen from Figure 9, this can be achieved by increasing \( F_{sw} \) from a late Cenozoic value of 0.2–0.3 – 0.35 or decreasing \( \alpha \) by 1%–1.5%. It can also be seen that a magmatic decarbonation flux of 50%–100% of the Cenozoic CO2 flux into the exogenic system (equivalent to increasing the total inputs of C into the exogenic system by 1.5–2 times) can achieve the same effect. While it is difficult to separate the contributions of all these variables from the seawater carbon isotopic record, this analysis shows that the addition of crustal carbonate decarbonation can lead to a significant positive shift in the baseline C isotopic composition of the exogenic system.

Geotherms and Metamorphic Reactions

The steady-state geotherm in Figure 2A was calculated from the heat diffusion equation assuming a 25 °C surface temperature, and a fixed surface heat flow (80 mW/m²) for a two-layer model (35 km crust overlying mantle). The crust has a heat production of 1.7 µW/m³ and the mantle heat production is assumed to be zero. This geotherm is meant as a possible reference state for the lithosphere in an active arc setting.

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