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Key Points:

- Waning volcanic degassing along the southern Eurasian margin is a possible cause of the long-term Cenozoic climate cooling
- A climate change-volcanism feedback during glacial-interglacial cycles explains the change in shape of late Cenozoic climate oscillations

Correspondence to:

P. Sternai,
pietro.sternai@unimib.it

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Magmatic Forcing of Cenozoic Climate?

Pietro Sternai², Luca Caricchi¹, Claudia Pasquero², Eduardo Garzanti², Douwe J. J. van Hinsbergen³, and Sébastien Castelltort¹

¹Department of Earth Sciences, University of Geneva, Geneva, Switzerland, ²Department of Earth and Environmental Sciences, University of Milano-Bicocca, Milan, Italy, ³Department of Earth Sciences, Utrecht University, Utrecht, Netherlands

Abstract Established theories ascribe much of the observed long-term Cenozoic climate cooling to atmospheric carbon consumption by erosion and weathering of tectonically uplifted terrains, but climatic effects due to changes in magmatism and carbon degassing are also involved. At timescales comparable to those of Milankovitch cycles, late Cenozoic building/melting of continental ice sheets, erosion, and sea level changes can affect magmatism, which provides an opportunity to explore possible feedbacks between climate and volcanic changes. Existing data show that extinction of Neo-Tethyan volcanic arcs is largely synchronous with phases of atmospheric carbon reduction, suggesting waning degassing as a possible contribution to climate cooling throughout the early to middle Cenozoic. In addition, the increase in atmospheric CO₂ concentrations during the last deglaciation may be ascribed to enhanced volcanism and carbon emissions due to unloading of active magmatic provinces on continents. The deglacial rise in atmospheric CO₂ points to a mutual feedback between climate and volcanism mediated by the redistribution of surface masses and carbon emissions. This may explain the progression to higher amplitude and increasingly asymmetric cycles of late Cenozoic climate oscillations. Unifying theories relating tectonic, erosional, climatic, and magmatic changes across timescales via the carbon cycle offer an opportunity for future research into the coupling between surface and deep Earth processes.

Plain Language Summary Among the most fascinating contemporary developments in the Earth Sciences is the idea that plate tectonics and climate changes are coupled through complex cycles. More than four decades of study have revealed tantalizing examples of relationships between processes near the Earth's surface and deeper within. Accounting for such a *surface-deep Earth process coupling* has improved our understanding of major climate and tectonic events for virtually all timescales. So far, however, the role of magmatism was overlooked. Magmatism exerts a strong influence on the amount of greenhouse gasses in the atmosphere because of volcanic outgassing. In turn, tectonic and climatic changes control the transfer of rocks, water, and ice masses at the Earth surface and within the Earth's interior, thereby affecting the production, transfer, and eruption of magmas. Unravelling the interactions between surface and deep Earth processes accounting for magmatism will help managing natural resources and mitigating natural hazards. Even more importantly, understanding how climate is naturally related to plate tectonics and magmatism will allow scientists to assess the anthropogenic perturbations to the Earth system and anticipate ongoing and future climate changes.

1. Rationale: Coupling Surface and Deep Earth Processes

Mountain ranges, sedimentary basins, subduction zones, volcanic arcs, oceanic ridges, and nearly all major geological features testify fluxes of geological materials occurring *at and across* the Earth's surface. Interactions between the solid Earth (i.e., our planet's solid surface and its interior) and the fluid Earth (i.e., our planet's wet or icy surface and the atmosphere) are thus ubiquitous and occur across timescales (e.g., Broecker, 2018). Correspondences between global climatic and tectonic changes throughout the Cenozoic fostered our understanding of the evolution of mountain ranges (e.g., Molnar & England, 1990), rock exhumation (Ruddiman, 1997), and the geological carbon cycle (e.g., Raymo & Ruddiman, 1992). The main logical link to relate past climate and tectonic changes is their synchronicity. However, limited temporal resolution in the geological archives prevents a clear recognition of the causative relationships behind such changes. Which mechanisms control the coupling between surface and deep Earth processes? Which are the characteristic timescales and magnitudes of the feedbacks involved? Answering these questions is still a challenge in the Earth Sciences (e.g., NAP, 2012; Frontiers, Media, 2015; Huntington et al.,

2017). A continuously growing body of observational constraints and advances in coupled processes modeling increases opportunities to progress in this research.

Here, we focus on the Cenozoic (last ~65 Myr), which includes several well-documented natural experiments reporting on the coupling between solid and fluid Earth processes (e.g., Miller et al., 1987; Zachos et al., 2008). In section 2, we address processes occurring at multimillion years timescales. First, we review an ongoing debate on causal relationships between the India-Eurasia convergence and collision history and climate cooling in the Neogene-Quaternary via erosion and weathering of tectonically uplifted terrains. We then develop the debate supporting the hypothesis that variations in volcanic degassing due to extinction of Neo-Tethyan volcanic arcs may have contributed to climate cooling since the Eocene. In section 3, we address the coupling between surface and deep processes at timescales comparable to those of Milankovitch cycles. Previous studies suggested that the waxing and waning of continental ice sheets, erosion, and sea level changes during climate oscillations affect the volcanic activity. Here, a simple model of radiative equilibrium is used to evaluate whether estimated variations of atmospheric CO₂ concentrations due to volcanic changes induced by the surface mass redistribution throughout glacial-interglacial cycles may affect the amplitude and symmetry of Pleistocene climate oscillations consistently with observations. In section 4, we outline open questions and opportunities for research focused on the coupling between surface and deep Earth processes. Quantifying feedbacks between climate and tectonics accounting for magmatism is a compelling challenge to progress our understanding of the coupling between surface and deep Earth processes, toward a holistic understanding of the functioning of the Earth system.

2. Solid Earth Control on Cenozoic Climate Cooling

The oxygen isotope composition of ocean sediments is sensitive to ocean temperature and ice volumes and shows unsteady but continuous increase in $\delta^{18}\text{O}$ values since ~50 Ma (Figure 1a). This implies long-term global cooling following the late Paleocene-early Eocene (~60–50 Ma) climate warming (e.g., Zachos et al., 2001). Antarctica was ice free until about ~30 Ma (e.g., Barrett et al., 1987; DeConto & Pollard, 2003; Pagani et al., 2011), and glaciation of the Northern Hemisphere only started during the Plio-Quaternary (last ~5 Myr; e.g., Raymo, 1994; Willeit et al., 2015). Thus, early Cenozoic $\delta^{18}\text{O}$ trends are driven primarily by temperature changes. The variability of the $\delta^{18}\text{O}$ record increased after ~30 Ma and became even larger in the last ~3 Myr, suggesting an increasingly dominant coupling with ice cover (e.g., Zachos et al., 2001; see section 3). Enhanced differentiation of the midlatitude continents into areas of wetter and drier climates is also inferred from the terrestrial fossil record and abundant glacial detritus blanketing middle-to-high latitudes worldwide (e.g., Ruddiman et al., 1989). For more than a century, scientists have been searching for explanations for the Cenozoic climate evolution (e.g., Broecker, 2018; Chamberlin, 1899), often alluding to a dominant, but still elusive, role of plate tectonics.

2.1. Tectonic Influence on the Ocean and Atmosphere Circulation Patterns

Paleogeographic changes, involving the formation and destruction of oceanic corridors and orographic barriers, are a straightforward mechanism through which plate tectonics may affect climate. Overall, global temperatures decrease when increasing portions of energy-absorbing oceans at the tropics are replaced by energy-reflective continental land (e.g., Crowley et al., 1987). However, continental migration toward the tropics over the past 100 Myr is limited and unlikely to explain Cenozoic cooling (e.g., Barron, 1985; Godd  ris et al., 2014). Thermal insulation due to the development of the circum-Antarctic currents, when Antarctica separated from Australia in the middle to late Oligocene (e.g., Barrett et al., 1987; Kennett et al., 1974), has also been invoked to explain synchronous cooling of Antarctica (e.g., DeConto & Pollard, 2003; Pagani et al., 2011). However, models suggest that this event would also reduce precipitation in the polar region, in turn inhibiting the onset of glaciation (e.g., Oglesby, 1989). A long-standing idea is that the emergence of the Isthmus of Panama in the late Pliocene may have triggered the onset of Northern Hemisphere glaciations (e.g., Bartoli et al., 2005; Haug & Tiedemann, 1998; Raymo, 1994). However, numerical simulations suggest that high-latitude cooling may also occur without the closure of the Panama gateway (e.g., Maier-Reimer et al., 1990; Murdock et al., 1997). An additional hypothesis is that Cenozoic true polar wander—the rotation of the Earth and mantle relative to the spin axis in response to change in solid Earth density distributions—may have preconditioned the northern Atlantic for glaciation (Steinberger et al., 2015).

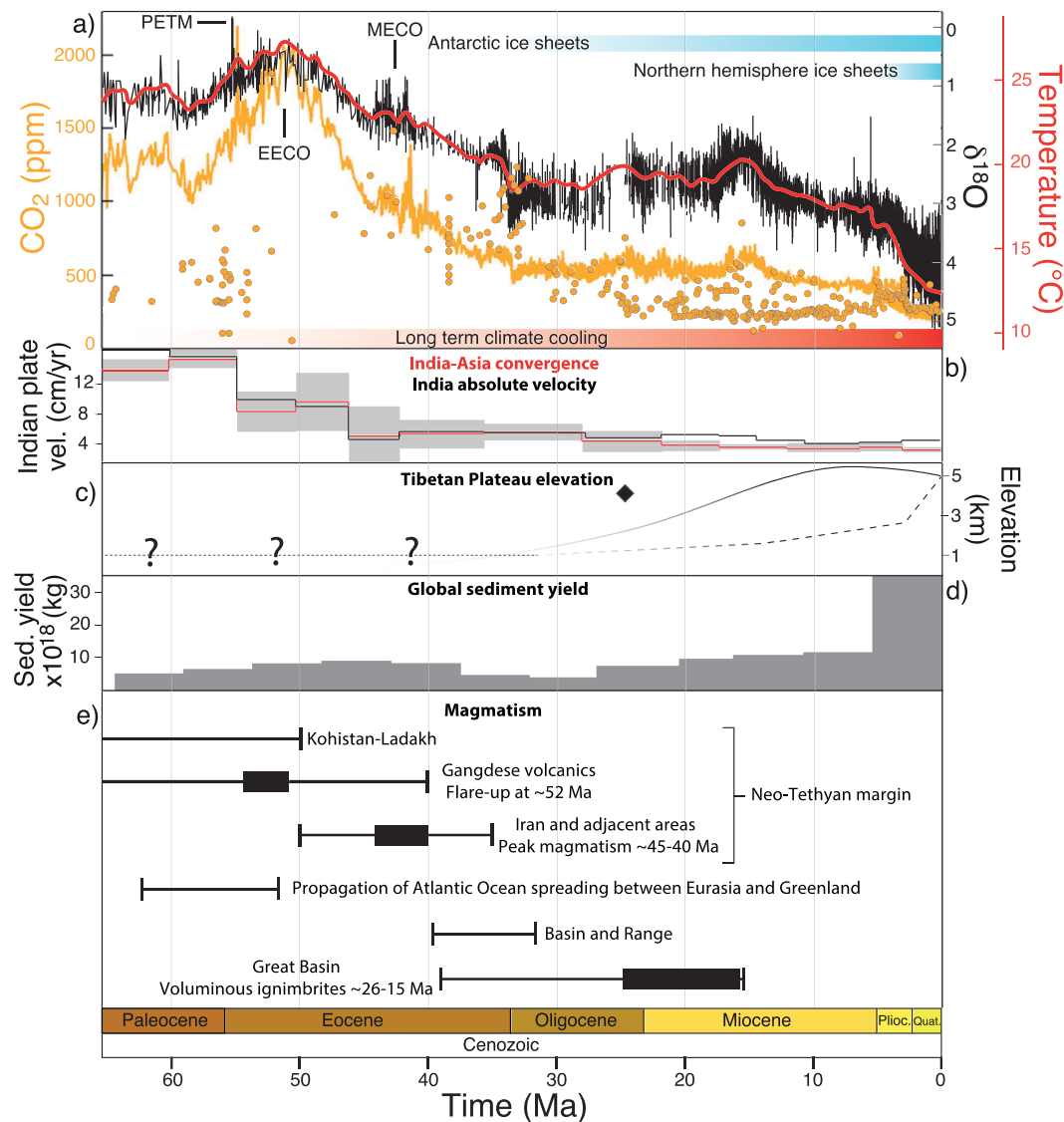


Figure 1. Compilation of datasets reporting on Cenozoic trends. (a) $\delta^{18}\text{O}$, surface temperature, and the atmospheric CO_2 concentration required to yield the global temperature change if climate sensitivity is $0.75^\circ\text{C}/\text{W}/\text{m}^2$ (Hansen et al., 2013; Zachos et al., 2001). Orange dots show the CO_2 proxy records as from Beerling and Royer (2011) and references therein (error bars are not shown for figure legibility). PETM, EECO and MECO stand for Paleocene-Eocene Thermal Maximum, Early-Eocene Climate Optimum and Middle-Eocene Climate Optimum, respectively. (b) India-Asia convergence (van Hinsbergen et al., 2011). (c) Mean Tibetan Plateau elevation; solid line from Fielding (1996), dashed line from Xu (1981), and diamond from DeCelles et al. (2007). Pre-Oligocene values (dotted line) are speculative (see text). (d) Global sediment yield (Peizhen et al., 2001). (e) Magmatic activity (nonexhaustive list; see also section 2.2.2 and references therein).

Climatic effects have also been linked to late Cenozoic growth of the Tibetan Plateau, the greatest topographic feature on Earth, which is inferred to have perturbed atmospheric circulation at the scale of the entire Northern Hemisphere (e.g., Ruddiman & Raymo, 1988; Ruddiman & Kutzbach, 1989; Molnar et al., 2010). Topography in the Tibetan Plateau was built up in several steps, likely starting well before the collision between India and Asia (Murphy et al., 1997; Wang et al., 2017). Paleoelevations may have been comparable to the present since at least the late Oligocene (DeCelles et al., 2007; Dupont-Nivet et al., 2008; Van Der Beek et al., 2009), but estimated plateau elevations during the Eocene vary from near sea level to ~5 km (e.g., Botsyun et al., 2019; Ding et al., 2014; Wei et al., 2016). Normal faulting in Tibet at about 13 Ma suggests an increase in Plateau elevation at that time (Molnar et al., 1993; Murphy et al., 2009) and may have had climatic effects such as initiation of aeolian accumulation in the main part of the Chinese Loess Plateau and the onset of intense monsoonal circulation (e.g., Sun et al., 2008, 2014). Numerical models indicate that a low-altitude Tibetan Plateau before the Eocene would have resulted in a trend from equable moist temperate

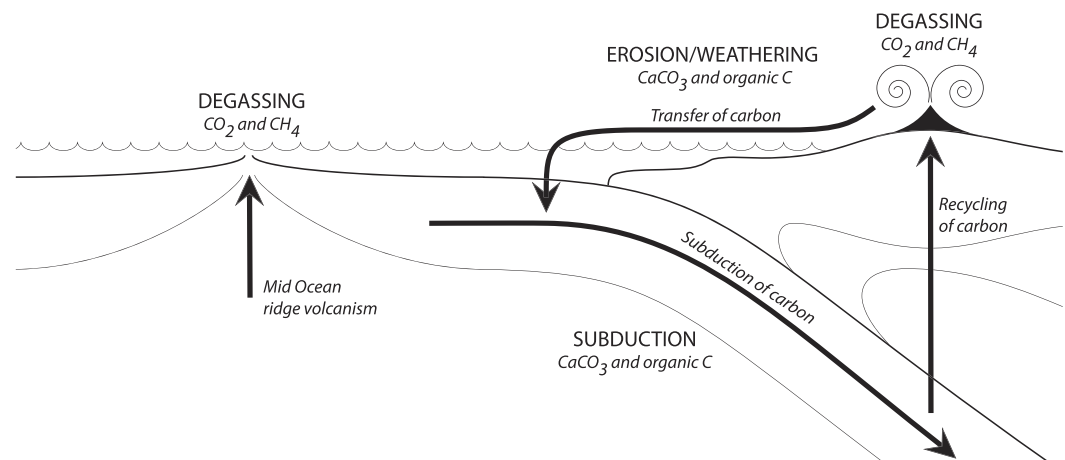


Figure 2. Schematic representation of sources and sinks of surface carbon.

climates in the early Cenozoic to increased seasonality and regional differentiation of climate in the Northern Hemisphere today (Ruddiman & Kutzbach, 1989), overall consistent with paleobotanical and paleoenvironmental records (e.g., Hren et al., 2009; Xu, 1981). However, predictions do not include a pronounced drop in high-latitude temperatures as the plateau elevations are increased, even when feedbacks (e.g., snow albedo, sea ice, and mixed-layer ocean temperatures) are taken into account. The development of high topography and/or the formation of marine sills or gateways alone thus seems insufficient to drive the long-term Cenozoic evolution. Additional factors such as changes in atmospheric or ocean composition are required (e.g., Raymo & Ruddiman, 1992).

2.2. Tectonic Influence on the Geological Carbon Cycle

At timescales of millions to tens of millions of years, the geological carbon cycle modulates the storage of carbon into rocks and the release of carbon into the ocean and atmosphere (e.g., Sundquist, 1985), thereby linking the evolution of climate and life to plate tectonics (Figure 2). Variations in the ocean and atmosphere concentration of radiative gases (particularly CO_2) can alter the Earth's climate. These express a dynamic equilibrium between sinks of carbon by chemical weathering of silicate minerals, precipitation/deposition of carbonates in the ocean and burial of organic matter in sediments (e.g., Galy et al., 2007), and sources of carbon by volcanic emissions, hydrothermal systems, and metamorphism (e.g., Walker et al., 1981; Berner & Lasaga, 1989; Berner, 2003). Relevant reactions can be simplified as



Geological proxies suggest decreasing atmospheric CO_2 levels since the Triassic followed by an early Cenozoic (~60–50 Ma) increase in CO_2 values (Figure 1a), including the Paleocene-Eocene Thermal Maximum at ~55.5 Ma and the Early-Eocene Climate Optimum at 52.9–50.7 Ma (e.g., Anagnostou et al., 2016; Beerling & Royer, 2011; Hansen et al., 2013; Zachos et al., 2001). Then, two subsequent phases of reduction of atmospheric CO_2 concentrations occurred. The first phase lasted until the Oligocene and included ~500 kyr of climate amelioration, the middle Eocene Climate Optimum, at ~40 Ma (e.g., Sluijs et al., 2013). The second phase began in the middle Miocene and continues today (although it is currently outpaced by short-term natural and anthropogenic emissions). Two end-member hypotheses have been formulated to explain the long-term Cenozoic cooling via a tectonic-controlled reduction in atmospheric carbon levels: Climate cooling may be driven either primarily by an increase in the global weathering rate (the main sink of surface carbon) on the one hand and by a decrease in the global degassing rate (the main source of surface carbon) on the other hand. We review these two hypotheses hereafter.

2.2.1. Increase in Global Weathering

Chemical weathering is a function of continental relief (e.g., Raymo et al., 1988; Ruddiman, 1997; Von Blanckenburg, 2006). During the Cenozoic, the convergence between India and Eurasia resulted in the deformation, uplift, and erosion of the Indian passive margin in the Himalayas, and in the formation of

the Tibetan Plateau in Southern Asia. In this region, incident solar heating in summer drives the strong atmospheric convection and rainfall associated with the Asian monsoon (e.g., Molnar et al., 1993). The presence of a large elevated plateau in the proximity of a warm ocean leads to very high weathering rates in the Himalayan region (Pinet & Souriau, 1988). Time constraints on the growth of the Tibetan Plateau to present-day elevations, however, are ambiguous (e.g., Botsyun et al., 2019; van Hinsbergen & Boschman, 2019; Figures 1b and 1c and section 2.1), which hampers establishing a straightforward correlation between climate cooling and the reduction of atmospheric CO₂ concentration due to Himalayan-Tibetan uplift and associated enhanced weathering. As an alternative or complementary mechanism, Jagoutz et al. (2016) propose atmospheric CO₂ drawdown due to weathering of mafic and ultramafic rocks between ~50 and 40 Ma (see also Macdonald et al., 2019).

The shift toward a cooler and stormier climate may have affected erosion also through a change from fluvial- to glacial-dominated conditions since the Plio-Quaternary (e.g., Molnar & England, 1990). The possible fourfold to fivefold increase in sediment deposition rates in most marine and oceanic basins since ~5 Ma (Figure 1d) would suggest increased erosion of most orogenic belts worldwide, ascribed to the onset of the Northern Hemisphere glaciation (e.g., Hay et al., 1988; Donnelly, 1982; Peizhen et al., 2001). Molnar and England (1990) further proposed that enhanced erosion and valley carving in the Plio-Quaternary would explain inferred synchronous uplift of mountain peaks worldwide (e.g., in the Alps, Pyrenees, and Tibet) through isostatic rebound of orogens following unloading. This idea provoked ongoing controversies because a global increase in mountain erosion should increase global weathering and CO₂ consumption (e.g., Riebe et al., 2004) and thus be the cause rather than the effect of climate cooling. On the one hand, the global increase in erosion and sedimentation may be related to a preservation bias in the sedimentary record (i.e., “Sadler effect”; Willenbring & von Blanckenburg, 2010; Willenbring & Jerolmack, 2016). On the other hand, the thermochronological record, which should be insensitive to preservation biases, corroborates the worldwide acceleration of mountain erosion (Herman et al., 2013). Although the analysis by Herman et al. (2013) was recently questioned (Schildgen et al., 2018), a number of geochemical proxies (e.g., Li and Sr isotopes) suggest a change in ocean composition during the late Cenozoic that may be related to increased worldwide erosion and weathering (e.g., Broecker, 2018; Edmond, 1992; Misra & Froelich, 2012). Assessing the links between mountain uplift, erosion, and weathering has been a major concern of research regarding the surface-deep Earth processes coupling. However, it remains unclear whether tectonics controls climate through mountain uplift, erosion, and weathering or rather if climate forces mountain uplift through erosion and isostatic rebound during the late Cenozoic. Despite a lack of consensus, more than four decades of intense debate suggest that a mutual feedback between tectonic uplift of mountain ranges and climate cooling via erosion and weathering does exist. Alternatively, or in addition, changes of global degassing rate following major tectonic events contributed to Cenozoic climate changes (e.g., Johnston et al., 2011; McKenzie et al., 2016; Brune et al., 2017; Godderis & Donnadieu, 2017), a possibility that is examined in more detail below.

2.2.2. Decrease in Global Degassing

Although climate cooling since the Triassic has been shown to correlate with changes in plate boundary magmatism (e.g., Li & Elderfield, 2013; Van Der Meer et al., 2014), the decrease in global volcanic degassing hypothesis received less attention compared to the increase in global weathering model of Cenozoic climate cooling. The overall temporal correspondence between waning of Neo-Tethyan volcanism and major climate deterioration throughout the Eocene and early Oligocene suggests a direct causal relationship between these observations. Yet, as we shall outline hereafter, other major geodynamic events occurred at roughly the same time and may have conditioned volcanic outgassing and the Cenozoic climate history. In addition, the extinction of volcanic arcs is inherently related to orogenic processes, which involve changes in erosion and weathering. Thus, the contributions of magmatic and weathering changes to the Cenozoic climate evolution are related. Our purpose here is to point out the potential climatic implications of magmatic changes throughout the Cenozoic, while constraining both contributions of magmatism and weathering to Cenozoic climate changes should be the objective of future research.

2.2.2.1. Neo-Tethys Closure and Associated Magmatism

Throughout the Mesozoic and into the Eocene, the Neo-Tethyan seaways extended east-west for more than 15,000 km (e.g., Seton et al., 2012). In the Late Cretaceous, the Neo-Tethys Ocean included a composite, east-west trending subduction system about 13,000 km long, with associated volcanic arcs along the

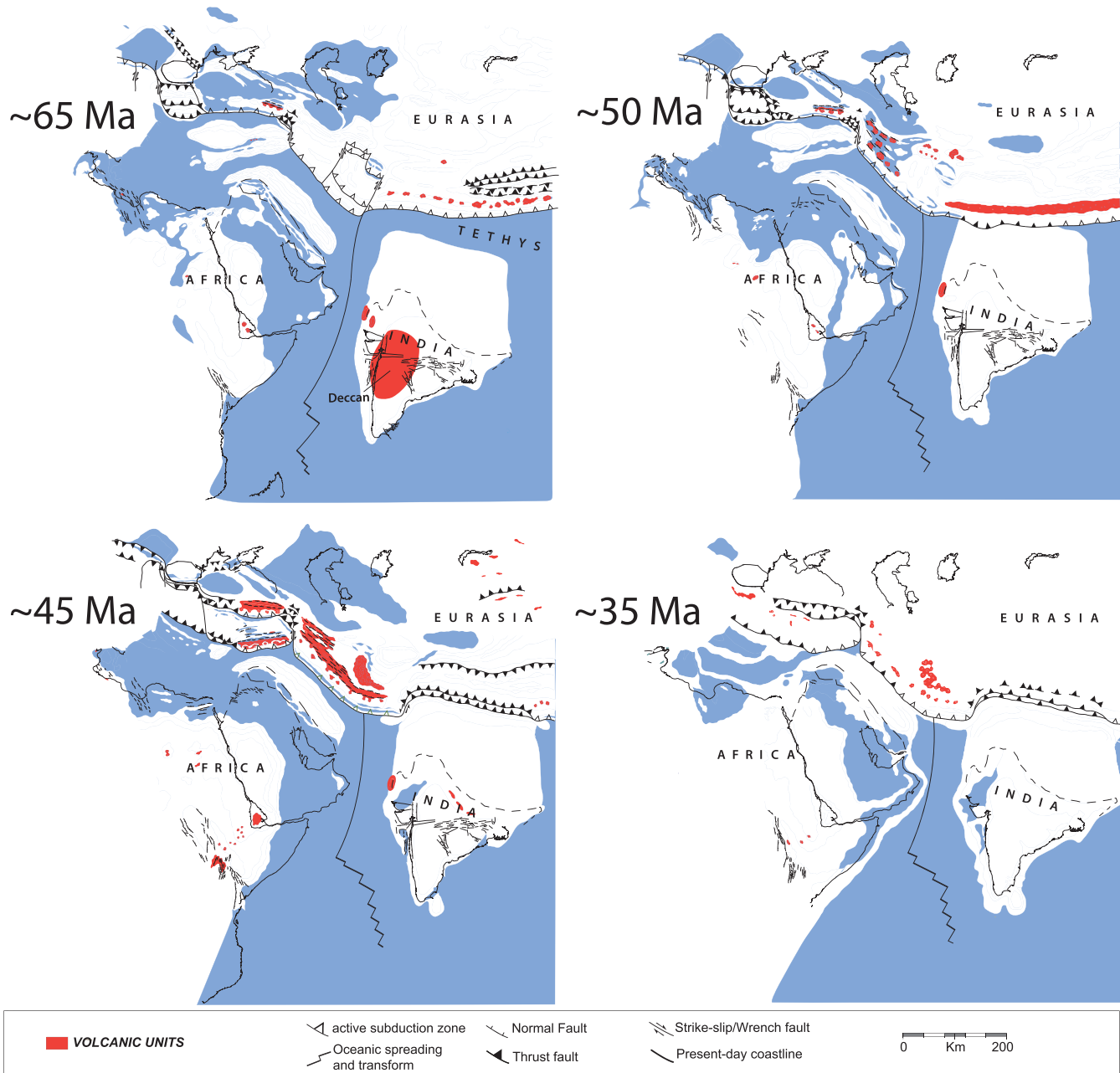


Figure 3. Paleotectonic maps of the Neo-Tethyan margin (modified after the paleotectonic reconstructions by the DARIUS program, 2018, <http://istep.dgs.jussieu.fr/darius/maps.html>).

southern continental margin of Eurasia and within the Neotethys ocean (e.g., Dercourt et al., 2000; Stampfli & Borel, 2002; Maffione et al., 2017; Guilmette et al., 2018). A double-subduction system is unequivocal between Arabia and Eurasia (e.g., Agard et al., 2007; van Hinsbergen, Maffione, et al., 2019) and was also proposed to have existed between India and Asia (e.g., Aitchison et al., 2007; Jagoutz et al., 2015), although the evidence there is debated (e.g., Hu et al., 2016; Huang et al., 2015). In any case, subduction of Indian plate lithosphere remained active until well after initial collision of the northernmost continental crust of the Indian plate with Asia (Figure 3). The timing of initial India-Asia collision is also controversial, with proposed ages from >60 and <35 Ma (e.g., Molnar & Tapponnier, 1975; Searle et al., 1987; Garzanti et al.,

1987; Aitchison et al., 2007; Bouilhol et al., 2013; Orme et al., 2015; Garzanti, 2008), partly depending on the assumed definition of continental collision (e.g., the end of marine sedimentation versus the end of ocean-continent subduction) or the method used to constrain the age of collision (e.g., stratigraphy, sedimentology, and style of deformation). The arrival of the edge of the Indian continental margin at the trench adjacent to southern Tibet (Beck et al., 1996) has been recently constrained as 59 ± 1 Ma (Decelles et al., 2014; Wu et al., 2014; Hu et al., 2015).

The climax of arc magmatism, however, trailed initial collision by some 7 Myr and is followed by progressive demise. Evidence of extensive Late Cretaceous to Paleogene volcanic activity is widespread along the southern margin of Eurasia (Figure 3). In southern Tibet, the Gangdese volcanic arc was voluminous until early Eocene times (e.g., Dewey et al., 1988; Treolar et al., 1996; Yin & Harrison, 2000) and recorded an ignimbrite flare-up stage around 52 Ma (e.g., Ji et al., 2009; Zhu et al., 2015), followed by return to background activity and slow extinction lasting until the late Eocene (e.g., Sanchez et al., 2013). The short-lived magmatic flare-up is in striking coincidence with the Early-Eocene Climate Optimum, suggesting a cause-effect relationship between the magmatic and climatic event, a possibility that was only marginally investigated thus far. Subduction of the carbonate-rich Indian passive margin succession in the middle Paleocene may have enhanced carbon recycling at the Transhimalayan volcanic arc (e.g., Kent & Muttoni, 2008; van Hinsbergen et al., 2019; Hoareau et al., 2015). Hoareau et al. (2015) modeled the volume of subducted materials and the amount of CO_2 emitted along the northern Tethyan margin and the effects of estimated CO_2 fluxes on global climate. Enhanced CO_2 degassing due to subduction of thicker Indian continental margin sediments after ~60 Ma would raise surface temperatures to some extent, consistently with the observed early Eocene climate warming. Similarly, Kerrick and Caldeira (1994) suggested that carbon degassing due to metamorphism of Indian margin rocks along the Himalayan belt enhanced paleoatmospheric carbon levels, although revised estimates point to a minor contribution to early Eocene warming (Kerrick & Caldeira, 1998). A reduction of carbon dioxide emissions is inferred between ~50 and 40 Ma, synchronous with the observed phase of climate cooling, when Tibetan arc volcanism was waning after collision between India and Eurasia (Hoareau et al., 2015; Jagoutz et al., 2016; Figure 1).

Between ~50 and ~35 Ma there has been the demise of a highly productive magmatic arc/back-arc system across much of Iran and adjacent areas (e.g., Agard et al., 2011; Schleiffarth et al., 2018). This magmatism was a net source of atmospheric CO_2 . Peak magmatism along the Arabia-Eurasia margin between ~45 and 40 Ma (e.g., Kazmin et al., 1986; Alpaslan et al., 2004; Arslan & Aslan, 2006; Vincent et al., 2005; Figure 1e) could have promoted a rise in atmospheric carbon concentrations during the middle Eocene climate optimum (Allen & Armstrong, 2008), which was previously attributed to an unspecified rise in igneous activity (Bohaty & Zachos, 2003). Only limited volcanic activity persisted into the Oligocene in Iran, though minor and sporadic magmatic events continue until the present (e.g., Pearce et al., 1990). Waning arc magmatism during the late Eocene in Iran and Anatolia is consistent with synchronous compressional deformation and topographic uplift (e.g., Allen & Armstrong, 2008; Gürer & van Hinsbergen, 2019; Vincent et al., 2005). The reduction in arc magmatism would have diminished carbon degassing into the atmosphere, and consequently global temperatures, consistently with observations (Figures 1e and 3).

2.2.2.2. Examples of Other Major Geodynamic Events and Associated Magmatic Changes

Obviously, this is not to dismiss the possible contribution of magmatic activity from igneous provinces other than the Southern Eurasian margin to early or middle Cenozoic climate changes. For instance, propagation of the Atlantic mid-ocean spreading between Eurasia and Greenland along the Reykjanes Ridge occurred by ~58 Ma (e.g., Seton et al., 2012; Figure 4) and the associated release of carbon likely contributed to the early Eocene climate amelioration. In the Caribbean region, there was continental underthrusting and ophiolite emplacement between 50 and 40 Ma on Cuba and Hispaniola (e.g., Boschman et al., 2014; Iturralde-Vinent et al., 2008), which may have contributed to waning global outgassing. Starting at ~39 Ma, large volumes of magmas were erupted from several volcanic centers in the Basin and Range, where magmatism ended by ~33 Ma (Gans, 1989), in remarkable synchronicity with the final phases of latest Eocene climate cooling. Ophiolite emplacement on Kamchatka and the Kuriles and the subsequent onset of subduction along the Aleutian and the Kuril trenches following arc-continent collision in Kamchatka and, farther south, along the Izu-Bonin-Marianas trench over a length comparable to that of the Neo-Tethyan margin occurred around ~50 Ma (e.g., Konstantinovskaia, 2001; Ishizuka et al., 2011; Domeier et al., 2017; Vaes et al., 2019; Figure 4). However, intraoceanic subduction zones commonly involve small amounts of carbon emissions

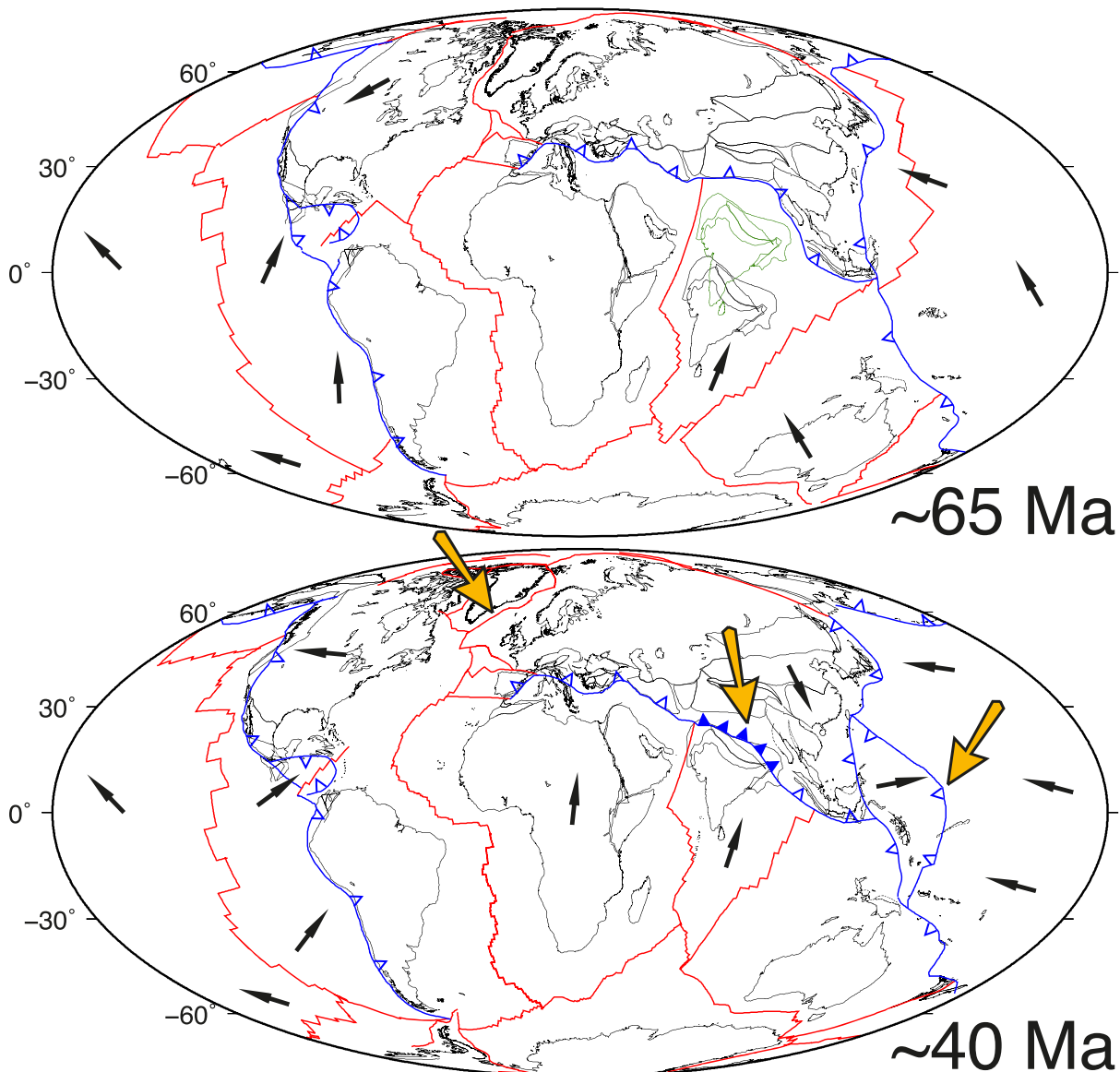


Figure 4. Global plate reconstruction at ~60 and ~40 Ma (modified after Seton et al., 2012). Blue and red lines show major subduction zones or thrusts and oceanic ridges or transform, respectively. The green line in the upper panel shows the position of India as suggested from the sedimentary and stratigraphic record (e.g., Hu et al., 2015), instead of the seafloor spreading history and global plate motion reconstruction (Seton et al., 2012). Yellow arrows point to major modifications of plate boundaries involving potentially significant magmatic changes: Demise of the Neo-Tethyan volcanic arc, onset of the Izu-Bonin-Marianas subduction, and propagation of ocean spreading between Greenland and Eurasia. Black arrows show the direction of motion of tectonic plates.

because few carbonate sediments are subducted (e.g., Johnston et al., 2011), either due to the distance from the continents or because intraoceanic trenches are typically deeper than the carbonate compensation depth. Tertiary volcanism in the Great Basin of western United States began in the Eocene with predominantly effusive and explosive volcanism and ended with late Oligocene and Miocene (~26–15 Ma) ignimbrite eruptions (Gans, 1987). Similarly, increased convergence rate normal to the Chilean Andes at ~25 Ma was followed by broadening of the volcanic arc into adjacent portions of Bolivia and Argentina (Pilger, 1984). The number of volcanic ash layers and volumetric estimates of igneous rocks erupted in the circum-Pacific region suggest increased volcanism in the Miocene (e.g., Donnelly, 1973; Kennett et al., 1977). These events may well have contributed to early and middle Miocene climate warming through enhanced carbon degassing (Kashiwagi & Shikazono, 2003). Although a number of previous global analyses show clear temporal correlations between volcanism and climate changes in the Paleozoic and Mesozoic (e.g., Lee et al.,

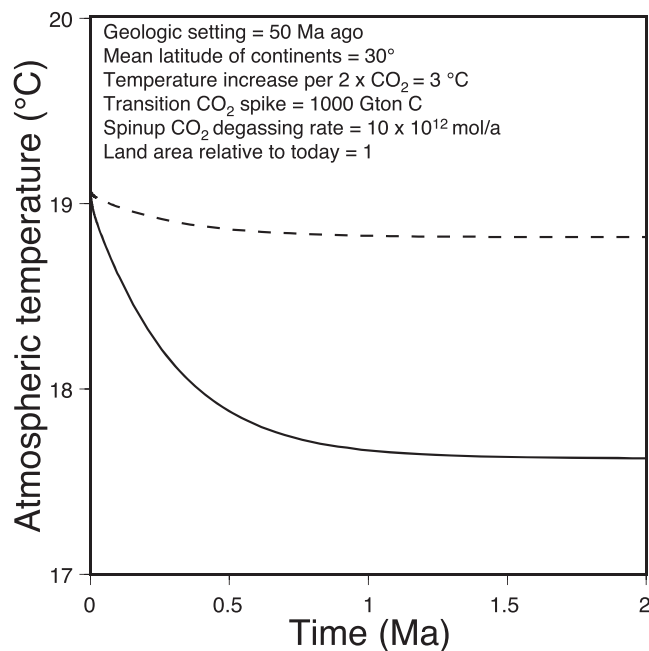


Figure 5. Estimated surface temperature change using the GEOCARB III model (<http://climatemodels.uchicago.edu/geocarb/>) imposing a decrease in global CO_2 degassing rate by 1% (dashed line) and 10% (solid line) from a spin-up CO_2 degassing rate of 10×10^{12} mol/a. The reader is referred to Berner and Kothavala (2001), specially Appendix 1, and Archer et al. (2009) for the relevant equations in GEOCARB III. Imposed boundary conditions are shown on the plot.

2013; McKenzie et al., 2016; Van Der Meer et al., 2014), these correlations have been barely tested for the entire Cenozoic.

2.3. Modeling Climatic Effects of Neo-Tethyan Arc Extinction

A global-scale analysis is required to assess thoroughly the potential magmatic control of Cenozoic climate changes. However, we emphasize here the role played by progressive extinction of Neo-Tethyan volcanic arcs during the Eocene (i.e., until ~35 Ma; Figures 1 and 3) and of the consequent reduction in global carbon emissions, which must have contributed at least to some extent to long-term Cenozoic climate cooling. To evaluate the climatic effects of the extinction of Neo-Tethyan volcanic arcs, it is necessary to translate fluxes of erupted magma into carbon fluxes. We have calibrated this relationship using the Indonesian arc as a modern analogue of the early Cenozoic Neo-Tethyan arc, because (i) it lies along the same plate margin, (ii) it shared a similar pre-Cenozoic and early Cenozoic tectonic and paleogeographic evolution, and (iii) it measures about the same length as the eastern Neo-Tethyan margin (pre-India-Eurasia collision) and western Neo-Tethyan margin (pre-Arabia-Eurasia collision; e.g., Ricou, 1994; Hall, 2002).

According to Burton et al. (2013, and references therein), estimated modern carbon emissions from the Indonesian volcanic arc contribute to global natural volcanic emissions by 1–10%. If these reference values are used as forcing terms for GEOCARB III models of the atmospheric $p\text{CO}_2$ and temperature evolution (Archer et al., 2009; Berner & Kothavala, 2001) and all other variables are kept constant, then predictions show cooling trends consistent in magnitude and timescale with that observed between, for example, 50 and 45 Ma (Figures 1 and 5). These first-order estimates are consistent with previous modeling efforts (e.g., Kashiwagi &

Shikazono, 2003) and suggest that a causal link between extinction of the Neo-Tethyan volcanic arcs and climate cooling throughout the Eocene-Oligocene is plausible.

We hope that these considerations may stimulate more thorough and systematic global-scale research efforts focused at constraining the relationships between magmatism and climate changes throughout the Cenozoic. We particularly stress that even though orogenic processes and the arrest of arc volcanism associated with continental subduction/collision are considered to produce effects on climate through enhanced erosion/weathering and reduced outgassing, respectively, the relationships between continental subduction/collision and magmatic activity are to date particularly poorly constrained.

3. Mutual Feedback Between Quaternary Climate and Volcanism

If climate changes are related to variations of the magmatic activity at long and intermediate timescales, then climate and magmatism may be even more tightly linked at short timescales such as those of Milankovitch cycles. Over the late Pliocene and Quaternary, benthic $\delta^{18}\text{O}$ values of deep-sea sediments and the Antarctic ice core records (see upper panels in Figure 6) show global temperature and ice volume oscillations, both expressions of glacial-interglacial cycles (e.g., Lisiecky & Raymo, 2007). Variations of insolation (i.e., solar energy received on the Earth's surface) and atmospheric CO_2 concentrations are primary drivers of climate oscillations and associated glacial-interglacial cycles (e.g., Bereiter et al., 2015; Imbrie, 1984). These show a puzzling progressive increase in amplitude and shift toward longer cooling than warming trends over the late Pleistocene (e.g., Lisiecky & Raymo, 2007). Because the parameters of the Earth's orbit vary within relatively stable boundaries at systematic Milankovitch periodicities and the solar radiation is constant at Myr timescales, it is safe to assume that the insolation curve maintained its amplitude and symmetry throughout at least the Quaternary (e.g., Broecker & Donk, 1970). Therefore, the change in shape of late Pleistocene climate oscillations must be driven by dynamics internal to the Earth system (e.g., Ashkenazy & Tziperman, 2004; Huybers, 2007; Liu & Herbert, 2004; Raymo, 1992). We hereby evaluate the possibility that such dynamics involve a feedback loop between glacial-interglacial cycles and volcanism via the

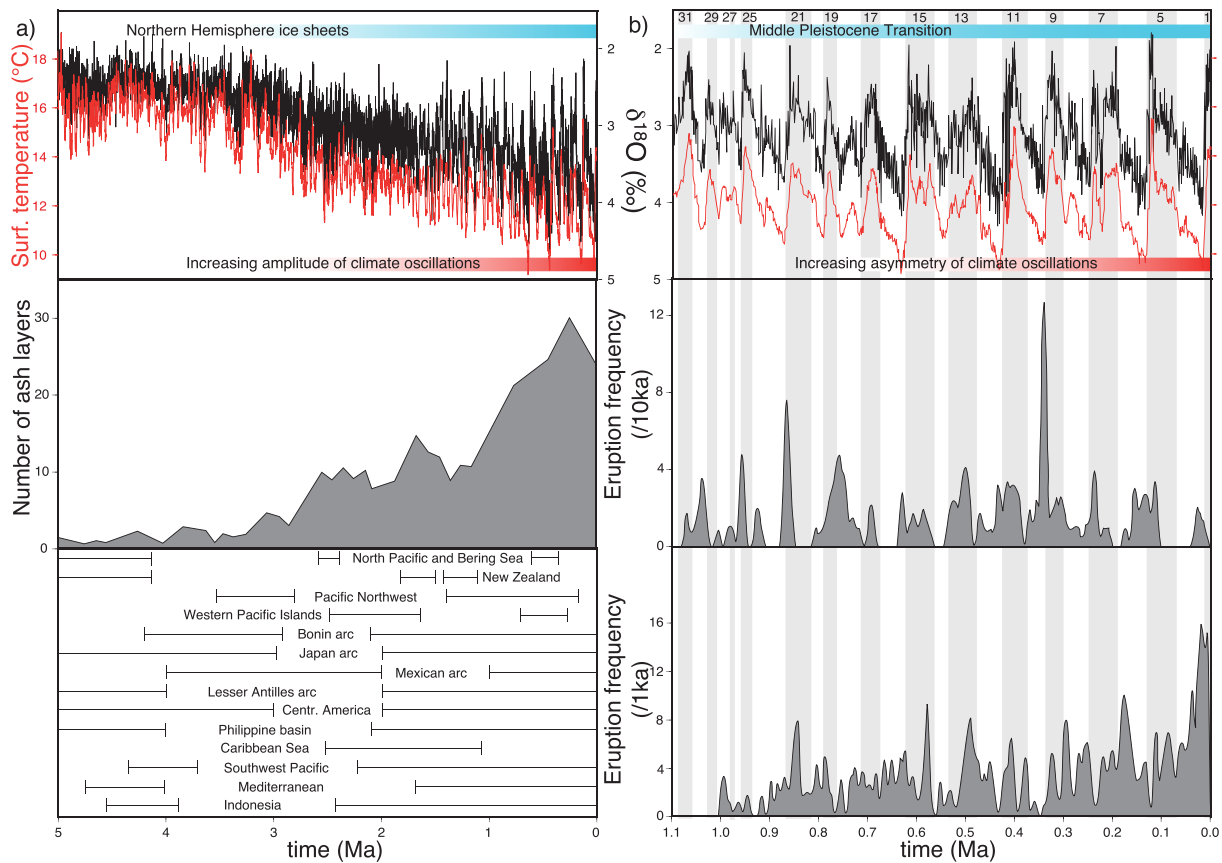


Figure 6. (a) Global $\delta^{18}\text{O}$ (Zachos et al., 2001) and temperature (Hansen et al., 2013) curves and tephra frequency in the Northwest Pacific drill (Prueher & Rea, 2001) over the last 5 Ma. The histogram is a three-point moving average after binning into 100 ka slots (see also Kutterolf et al., 2019). Periods of increased volcanic activity around the Pacific Ocean are shown in the lower panel (Kennett & Thunell, 1975; Kennett et al., 1977; Hein et al., 1978; Cambray & Cadet, 1994; Shane et al., 1995; Carey & Sigurdsson, 2000). (b) Global $\delta^{18}\text{O}$ curve (Zachos et al., 2001) and tephra frequency at the Izu Bonin Mariana arc (central panel, using 10 ka binning after Schindlbeck et al., 2018) and the Pacific Ring of Fire (lower panel, using 1 ka binning after Kutterolf et al., 2013). The Middle Pleistocene Transition leads from dominant ~40 ka periodicity of climate oscillations to dominant ~100 ka cycles. Vertical gray bars mark marine isotope stages (MIS) after Railsback et al. (2015).

building/melting of ice sheets, erosion, and volcanic CO_2 degassing that affected the shape of late Pleistocene climate oscillations, as previously hypothesized by, for example, Huybers and Langmuir (2009).

3.1. Glacial-Interglacial Cycles and the Forcing on Volcanism

In the past few decades, it has been proposed that climate changes may affect volcanic activity (Kutterolf et al., 2019, and references therein). Studies of several glaciated regions worldwide (e.g., Sigvaldason et al., 1992; MacLennan et al., 2002; Nowell et al., 2006; Rawson et al., 2015; Praetories et al., 2016) showed a significant increase in subaerial volcanic activity during postglacial warming. Huybers and Langmuir (2009) analyzed a global record of subaerial volcanic eruptions back to the last glacial period and showed an increase in eruption frequency of at least 50% during deglaciation. This correspondence is classically explained by a deglacial-triggering hypothesis (Hall, 1982), according to which the removal of ice caps during interglacials and associated continental lithospheric unloading leads to increased subaerial volcanic activity by reducing the confining pressure within crustal magma chambers (e.g., Sigvaldason et al., 1992; Jellinek et al., 2004; Pagli & Sigmundsson, 2008) and enhancing mantle melting (e.g., Jull & McKenzie, 1996; Tuffen, 2010). The modeling of Sternai et al. (2016) further indicates that erosion changes during glacial-interglacial cycles may produce similar unloading effects on magmatism as the melting of continental ice sheets.

Several authors identified relationships between volcanic cycles and orbital forcing at ~23, ~41, and ~100 kyr (Milankovitch, cycles). Paterne et al. (1990) recognized five periods of enhanced volcanic activity of the

Campanian province (Italy) with an ~ 23 kyr periodicity, correlating also with compositional changes in volcanic products. McGuire et al. (1997) observed enhanced tephra accumulation in the Mediterranean every ~ 22 kyr throughout the last ~ 80 kyr, correlating with the rate of sea level change. Glazner et al. (1999) investigated the distribution of intraplate volcanic events in eastern California over the last 800 kyr and found a correlation with glacial maxima. Jellinek et al. (2004) found an ~ 40 kyr periodicity (and more ambiguous ~ 17 and 23 kyr periodicities) in the first 400 kyr of the same data set as that of Glazner et al. (1999) and showed that periods of fast ice volume reduction anticipate volcanic pulses by ~ 3 and ~ 11 kyr for silicic and basaltic magmas, respectively. Schindlbeck et al. (2018) analyzed the ~ 1.1 Myr long tephra record of the Izu-Bonin arc (within the IODP Hole U1437B) and showed statistically significant spectral peaks with ~ 100 kyr periodicity (Figure 6), occurring systematically a few thousands of years after glacial maxima over the last ~ 0.7 Myr. Kutterolf et al. (2013) compiled a time history of volcanism from around the Pacific Ring of Fire, which accounts for about half of the global length of active plate subduction, and recognized an ~ 41 kyr periodicity, a correlation with the highest rate of decreasing global ice volumes, and a time lag between deglaciation and volcanic peaks of ~ 4 kyr. Further large-scale magmatic effects are observed in response to major modifications of surface conditions. Cambray et al. (1995) and Prueher and Rea (1998, 2001) emphasize the synchronism between the onset of the Northern Hemisphere glaciation at ~ 2.5 Ma and intense volcanic activity all around the North Pacific Ocean possibly due to sea level lowering (Figure 6), a link that was already hypothesized by Kennett and Thunell (1975). Sternai et al. (2017) proposed a link between sea level drop during the Messinian salinity crisis and a pulse of magmatic events in Mediterranean igneous provinces.

Melting ice sheets on land has the downstream consequence of raising sea level. This would imply that submarine volcanism may be inhibited by a deeper water column at the same time as subaerial volcanic eruptions are enhanced by glacial unloading on land. The seafloor bathymetry, primarily determined by oceanic crustal thickness, suggests a sensitivity of mid-ocean ridge volcanism to sea level changes on Milankovitch timescales (e.g., Crowley et al., 2015; Lund & Asimow, 2011; Schindlbeck et al., 2018; Tolstoy, 2015). A peak of hydrothermal activity ~ 15 ka after the last glacial maximum on the Mid-Atlantic Ridge (Middleton et al., 2016), Juan de Fuca Ridge (Costa et al., 2017), and at the East Pacific Rise (Lund et al., 2016) has been interpreted as the delayed melt formation following the hydrostatic pressure minimum.

3.2. Volcanic Forcing on the Shape of Glacial-Interglacial Cycles

The solid Earth vents CO_2 into the atmosphere through volcanism and magmatic processes, which appear to vary in response to the redistribution of the surface ice, water, and rock masses during glacial-interglacial cycles. Because changes in CO_2 removal rate by chemical weathering have a negligible effect on timescales comparable to those of glacial-interglacial cycles (e.g., Le Hir et al., 2009; Colbourn et al., 2015), the following question arises: Can variations of global CO_2 volcanic degassing driven by glacial-interglacial cycles affect the shape of climate oscillations? Since glaciation and deglaciation produce anticorrelated variations in subaerial and submarine volcanism (see section 3.1), the relative magnitude of these effects has to be quantified to answer the question.

Assuming that emissions are proportional to the frequency of eruptions, Huybers and Langmuir (2009) estimate the global evolution of volcanic CO_2 emissions since 40 ka (Figure 7a). Their model indicates an atmospheric CO_2 reduction by 5–20 ppm during the last glacial period (~ 40 – 18 ka), marginally consistent with the observed ~ 20 ppm decrease in atmospheric CO_2 data from the Dome C and Taylor Dome Antarctic ice cores (Indermühle et al., 2000; Monnin et al., 2004). During the initial phase of deglaciation (~ 18 – 13 ka) their model suggests a 5–40 ppm increase in CO_2 concentration, whereas observations show an ~ 50 ppm increase. During advanced deglaciation (~ 13 – 7 ka), the reconstruction indicates a 15–70 ppm increase in volcanic CO_2 , consistent with the observed ~ 40 ppm increase. According to their analysis, equilibration with oceanic volcanism compensates for up to $\sim 20\%$ of the increase in subaerial volcanic CO_2 flux during the deglaciation. The overall consistency between estimated volcanic emissions by Huybers and Langmuir (2009) and observed atmospheric concentrations of CO_2 suggests that volcanic changes may indeed be primary drivers of the glacial-interglacial climate variability. Huybers and Langmuir (2009) construct their time history of CO_2 fluxes using eruption frequencies rather than unloading histories, and so the potential contribution of erosion to increased volcanic degassing during the deglaciation (Sternai et al., 2016) is already accounted for. However, the contribution of erosion changes to unloading of active magmatic provinces is not acknowledged in the classic deglacial-triggering hypothesis of volcanic events (e.g., Hall, 1982; Sigvaldason et al., 1992; Jull &

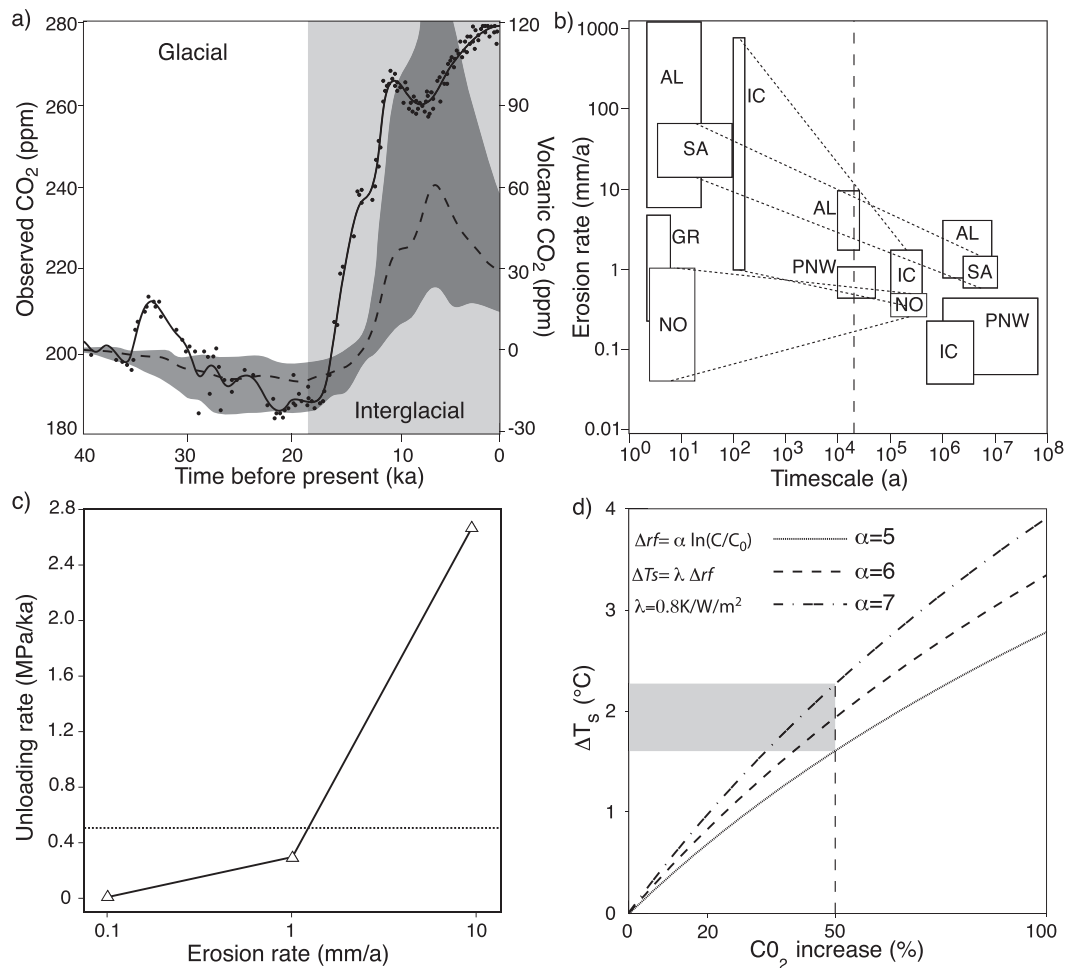


Figure 7. (a) CO₂ concentrations from the Dome C and Taylor Dome Antarctic ice cores 4 (dots), and the data smoothed over a 2 ka window (solid line) follow the left axis. Modeling results by Huybers and Langmuir (2009) constraining the contribution to atmospheric CO₂ from volcanic activity (dashed line) follow the right axis. The dark and light gray shaded regions represent the 90% confidence interval on the modeling estimates and the interglacial period, respectively. (b) Boxes represent ranges of erosion rates from glaciated catchments or proximal basins including errors in estimations (vertical) and resolved timescale (horizontal; Brandon et al., 1988; Hallet et al., 1996; Sheaf et al., 2003; Reiners et al., 2003; Koppes et al., 2006; Hebbeln et al., 2007; Geirsdottir et al., 2007; Berger & Spotila, 2008; Koppes et al., 2009; Cowton et al., 2012; Herman et al., 2013; Herman & Brandon, 2015). AL, Alaska; SA, Southern Andes; GR, Greenland; NO, Norway; PNW, Pacific Northwest; IC, Iceland. The vertical dashed line represents the approximate time since the LGM (Lisiecky & Raymo, 2007). (c) Estimated continental unloading rate owing to constant melting of 1,000 m of ice (dotted line for reference) and erosion (assuming a surface rock density of 2,700 kg/m³) throughout the interglacial. (d) Equilibrium change in global mean surface air temperature resulting from increasing atmospheric CO₂ concentrations (see text for detail).

McKenzie, 1996; Jellinek et al., 2004; Pagli & Sigmundsson, 2008; Tuffen, 2010). It thus seems important to remark here that given the density contrast between ice and surface rocks, erosion rate changes by ~1 mm/a or higher during deglaciation unload continents by a similar or greater amount than the melting of ice sheets (Sternai et al., 2016). Long-term proxies of global sediment efflux from mountainous regions show just such erosion rate variability (e.g., Peizhen et al., 2001; Koppes & Montgomery, 2009; Herman et al., 2013; Figures 7b and 7c). At secular to millennial timescales, erosional fluxes may be even higher (e.g., Koppes & Montgomery, 2009) and subject to strong variations due to modifications of the subglacial hydraulic system and water supply (e.g., Andrews et al., 2014; Flowers & Clarke, 2002; Herman et al., 2011; Sternai et al., 2013). The abrupt and high-magnitude magmatic pulses involved by such erosional changes are likely to force centennial- to millennial-scale variations of atmospheric greenhouse gases seemingly unrelated to ocean dynamics (e.g., Marcott et al., 2014). In addition, while reduced submarine volcanism due to sea level rise during interglacials buffers increased subaerial CO₂ emissions, the erosional forcing on subaerial emissions is unbalanced because the downstream effect of erosion is depositional loading of epicontinental or abyssal marine basins mostly located away from oceanic ridges.

It thus seems plausible that enhanced subaerial volcanic degassing due to combined ice melting and erosion may account for all of the observed ~ 90 ppm increase in atmospheric CO_2 concentration during the deglaciation, that is, an increase of $\sim 50\%$ with respect to glacial concentrations (Figures 7a and 7d). Arrhenius law may be used to calculate the resulting radiative forcing, rf , so that $rf = \alpha \cdot \ln \frac{C}{C_0}$ where α is a constant and $\frac{C}{C_0}$ is the relative increase in CO_2 concentration. The equilibrium change in global mean near-surface air temperature, T_s , is then calculated as $T_s = \lambda \cdot rf$, where λ is the climate sensitivity. Assuming λ equal to 0.8 K/W/m^2 (e.g., Royer et al., 2007), a T_s of about 2°C solely due to volcanic changes throughout glacial-interglacial cycles should be expected (Figure 7d). This volcanic forcing of temperature oscillations is concordant and largely in phase with that related to variations of the Earth's albedo during formation and melting of continental ice caps, producing an additional gain of surface temperatures through glacial-interglacial cycles of comparable magnitudes (e.g., Broccoli & Manabe, 1987; Dickinson et al., 1987; Meehl & Washington, 1990). A gain of temperature oscillations by up to $\sim 4^\circ\text{C}$ during the late Pleistocene with respect to prior the onset of the Northern Hemisphere glaciation is consistent with observations (Hansen et al., 2013; see top panel in Figure 6a). An increase in the amplitude of temperature oscillations is also observed after the Eocene-Oligocene boundary, when the Antarctic ice sheet is established (Figure 1a). The mutual influence between volcanism and climate oscillations in the presence of continental ice sheets, amplified by albedo effects and larger continental ice volumes in the Pleistocene, may thus represent a primary driver of the late Cenozoic progression toward higher-amplitude climate oscillations.

The climate-volcanism feedback during glacial-interglacial cycles would also generate the asymmetric cooling and warming trends of late Pleistocene climate oscillations (e.g., Lisiecky & Raymo, 2007). The main logic is that inhibition of subaerial eruptions during glacial periods forces accumulation of gasses in magmatic reservoirs, which are then released over a few thousands of years during the early interglacials (e.g., Jellinek et al., 2004). Assuming that the long-term weathering CO_2 sink is at equilibrium with the steady state volcanic CO_2 outgassing, the weathering carbon sink slightly dominates over inhibited volcanic carbon emissions when ice sheets grow, leading to a temporary reduction of atmospheric CO_2 , which sustains climate cooling. As soon as the orbital forcing of solar radiation overtakes the threshold to trigger the deglaciation, enhanced volcanic carbon outgassing dominates over the weathering carbon sink, in turn fostering climate warming and bringing the overall atmospheric CO_2 budget back to equilibrium. If, for instance, enhanced outgassing is extinguished in ~ 10 ka, then the cooling phase is forced to ~ 30 ka for early Pleistocene 40 ka cycles and to ~ 90 ka for late Pleistocene 100 ka cycles. The ~ 10 ka time window is chosen arbitrarily, but the duration of warming of late Pleistocene climate oscillations and the expected response time of magmatic systems to surface load changes constrain this value (e.g., Jellinek et al., 2004; Lisiecky & Raymo, 2007). Because both constraints are largely independent on the period of climate oscillations (i.e., 40 or 100 ka), fixing the duration of the phase of enhanced outgassing allows us to speculate that the longer the period of climate oscillations, the longer the phase of gas accumulation in magmatic reservoirs, the larger the outgassing rate during the phase of enhanced emissions, and the more pronounced the asymmetry between warming and cooling trends (Figure 8). This is consistent with the increasing amplitude and asymmetry of climate oscillations throughout the mid-Pleistocene transition from dominant 40 to 100 ka cycles (e.g., Lisiecky & Raymo, 2007). We hope that this statement may trigger and motivate further research on the volcano-climatic feedback during glacial-interglacial cycles.

4. Surface-Deep Earth Processes Coupling: Challenges and Opportunities

Once plate tectonics theory became established (e.g., Le Pichon, 1968; McKenzie & Parker, 1967), the correspondences between major geodynamic, magmatic, and climatic changes were readily recognized (e.g., Kennett & Thunell, 1975; Kennett et al., 1977; Ziegler et al., 1979; Raymo & Ruddiman, 1992; McKenzie et al., 2016). Assessing the physical mechanisms and chemical reactions that allow for the surface-deep Earth processes coupling has been a major concern of research ever since. Scientists focused on processes such as mountain building, erosion, and the waxing and waning of continental ice sheets. Disassembling a complex system of interactions into simpler relationships has been the main approach to progress on this research, and a broad range of geological data sets were used to develop theories relating mountain building and erosion, and next erosion and atmospheric CO_2 consumption. However, at the multimillion years time-scale, besides erosion and weathering of uplifted terrains, mountain building following continental

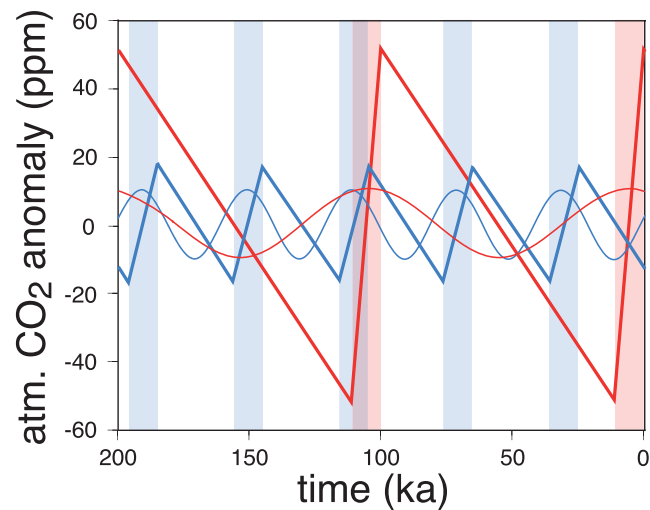


Figure 8. Thick lines show modeled atmospheric CO₂ variability associated with normalized sinusoidal radiative forcing at 40 kyr (thin blue line) and 100 kyr (thin red line). Shaded bars indicate the phase of enhanced outgassing during the deglaciation lasting for ~10 kyr (see text for detail). Ocean and atmospheric CO₂ are assumed to be in equilibrium, as their equilibration timescale is much shorter than the timescales considered here (Archer, 2005). Weathering is the only CO₂ sink considered and the weathering rate, \dot{W} , is assumed to be constant and equal to 0.1 Gt/a (Gaillardet et al., 1999; Moon et al., 2014). The volcanic CO₂ emission rate is assigned a glacial value, \dot{e} , and an enhanced value, \dot{E} , during the deglaciation calculated as $\dot{E}=2\dot{e}$ (Huybers & Langmuir, 2009). We also assume that the CO₂ budget is balanced over an entire climate cycle, so that $\dot{W}(g+d) = \dot{E}d + \dot{e}g$, where g and d are the duration of the glacial and deglacial periods, respectively. Thus, for 100 kyr climate cycles with deglaciation lasting for 10 kyr (see text for detail), $\dot{e}=0.09$ Gt/a and $\dot{E}=0.18$ Gt/a. For 40 kyr climate cycles with deglaciation lasting for 10 kyr, we adopt $\dot{e}=0.09$ Gt/a, leading to $\dot{E}=0.13$ Gt/a.

collision/subduction involves the closure of oceanic seaways and is typically followed, after a certain time lapse, by extinction of volcanic arcs, which also impact on climate. Early Cenozoic climate cooling is widely ascribed to atmospheric carbon consumption due to enhanced erosion and weathering following initiation of Himalayan growth, but Paleocene closure of Neo-Tethys and final waning of arc magmatism are currently unconstrained factors. In addition, the ~10 Myr long phase of climate warming preceding the onset of the Cenozoic cooling trend cannot be readily explained by theories referring only to enhanced weathering following topographic uplift because enhanced weathering involves a reduction of atmospheric CO₂ and thus climate cooling. Similarly, known effects of surface processes on magmatism and of magmatism on climate have not been duly considered to assess whether enhanced erosion following mountain building led to the onset of glaciation in the late Cenozoic or the onset of glaciation enhanced erosion rates and lithospheric unloading, thereby increasing mountain uplift through isostatic adjustment. Until the magmatic forcing of Cenozoic climate is quantified, the contribution of erosion/weathering of uplifted terrains to climate cooling will remain elusive. Continental collision/subduction and associated orogenesis increase weathering and cause arc demise, and both processes work in tandem to reduce the atmospheric carbon budget. Discerning between the two contributions to the Cenozoic climate evolution as indicated by paleoclimate proxies is a major challenge for future research. The links between surface mass redistribution and volcanism characterized by recent research focused on Plio-Quaternary glacial-interglacial cycles seem to offer an appropriate starting point for the development of a unifying theory relating tectonics, surface processes, magmatic, and climatic changes. In this way, we may eventually capture the essence of the surface-deep Earth processes coupling and thus gain new insights into the functioning of the Earth system (Figure 9).

The development and validation of such a theory involves a range of disciplines of the Earth Sciences and will require the conjugation and systematic analyses of global data and modeling. Besides extending the body of observational constraints reporting on the Cenozoic tectonic and sedimentary fluxes, improving data sets on the timing, magnitude, and chemical composition of major magmatic events is needed. Interpreting available data will require implementation of current models to account for phase transitions and/or multi-phase flow in order to investigate the coevolution of geodynamic and magmatic processes. The continuously

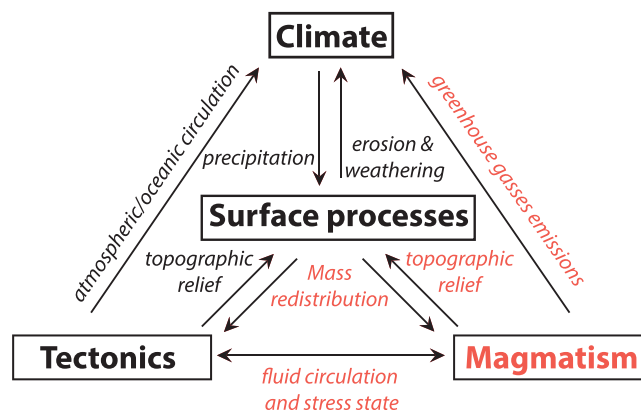


Figure 9. Schematic representation of the surface-deep Earth processes coupling. Assessing the role of magmatism in setting Cenozoic climate changes is one of the Grand Challenges for future research on the surface-deep Earth processes coupling.

increasing amount of experimental constraints on rock rheology and petrology and recent improvements in numerical techniques allowed the development of petrothermomechanical numerical models capable of handling intense deformation and magmatism (e.g., Bouilhol et al., 2015; Gerya & Yuen, 2007; Menant et al., 2016; Roche et al., 2018). A component of surface processes is often included into recent geodynamic numerical models (e.g., Ueda et al., 2015) and can benefit from a wide range of continuously growing geomorphological, sedimentological, biostratigraphical, and geophysical observations. Climate models are also continuously improved in order to account for multiple factors pertaining to the solid Earth such as variations of the tectonic or sedimentary fluxes (e.g., Archer et al., 2009; Zeebe, 2012). When boundary conditions cannot be constrained by direct observations, estimates stemming from independent petrothermomechanical and/or surface processes modeling results may be useful. Although challenging, assessing first-order relationships between major tectonic and climatic changes accounting for magmatism seems feasible as of now.

To what extent did the extinction of Neo-Tethyan volcanic arcs and/or other major geodynamic events affect global carbon degassing throughout the Cenozoic? Would associated changes in surface CO_2 concentrations involve a climatic response consistent with paleoclimate proxies? Understanding the post-50 Ma climate cooling may first require understanding the ~60–50 Ma climate warming, seemingly unrelated to weathering changes and suggesting a prominent magmatic forcing on climate. To this aim, we particularly stress the need for accurate and systematic reconstructions of plate kinematics and magmatic activity at a global scale to be used as observational constraints for modeling the associated perturbations of the geological carbon cycles. To what extent can the variability of atmospheric carbon during glacial-interglacial cycles be ascribed to changes in volcanic activity? Do glaciation and associated redistribution of surface masses act as phase locker between volcanic and climatic changes through a modulation of carbon emissions? Do anticorrelated variations of submarine and subaerial volcanism throughout glacial-interglacial cycles add up to positive or negative perturbations of the global long-term carbon degassing and how would these perturbations modulate the shape of Late Quaternary climate oscillations? Understanding the relationships between Plio-Quaternary climate oscillations and volcanism requires accurate quantifications of the surface ice water and rock mass redistribution associated with glaciation and in-depth assessments of the response of magmatic systems to perturbations of the surface loads. In general, any contribution that improves our understanding of carbon recycling at plate margins, volcanic arc extinction, surface load variations during glacial-interglacial cycles, and the carbon emission budget of submarine and subaerial volcanism will help answering these and other outstanding questions regarding the coupling between surface-Earth and deep-Earth processes.

Assessing the climatic response to long- and short-term magmatic forcing throughout the Cenozoic will not only be of scientific relevance but also produce invaluable societal implications. For instance, early Cenozoic climate changes provide insights into the coupling of climate and the carbon cycle and thus may help to predict the consequences of current carbon emissions in the future (e.g., Zachos et al., 2008). In addition, glaciers and ice sheets are a primary resource of fresh water and are often used in critical ways, while geothermal power plants around volcanic contexts worldwide should produce about 140 GWe (~8.3% of total electricity production) by 2050 (e.g., Bertani, 2016). Because enforced deglaciation since the industrial era may lead to increased volcanic activity and hazard (e.g., Pagli & Sigmundsson, 2008; Tuffen, 2010), understanding the natural joint variability of glaciated and volcanic environments is an urgent action to undertake in order to improve our ability to manage natural resources and hazards in fast-evolving and high-potential/risk terrains.

5. Boxed Text

The geological carbon cycle regulates the storage of CO_2 into rocks and the release of CO_2 into the ocean and atmosphere via weathering of surface minerals and volcanic emissions, thereby linking the evolution of

climate and life to plate tectonics. The last ~60 Myr are characterized by a long-term climate cooling trend, which may be ascribed to a combination of reduced atmospheric CO₂ due to increased global weathering and/or progressive reduction of CO₂ emissions following extinction of volcanic arcs. Assessing the magmatic contribution to the long-term climate evolution is particularly challenging because our knowledge regarding volcanic carbon emission and the timing and modalities of volcanic arc generation and extinction is still flawed. Major research efforts are thus required and should involve global-scale analyses of available data and the development of new observational constraints and models.

Climate cooling throughout the Cenozoic Era led to the onset of Northern Hemisphere glaciation in the Plio-Quaternary (~5 Ma to present). Because volcanic emissions affect climate, but the redistribution of surface ice, water, and rock masses associated with glacial-interglacial cycles affects volcanic activity, a feedback loop between volcanic and climatic change does exist. Paleoclimate proxies show a progressive increase of the amplitude and asymmetry of climate oscillations throughout the Late Quaternary, and our preliminary analysis suggests that the mutual feedback between volcanism and climate during icehouse stages may explain such observations. However, more thorough quantifications of the surface mass redistribution during glacial-interglacial cycles and a better understanding of the sensitivity of volcanic systems to surface load variations are required to assess the variability of subaerial and submarine volcanic carbon degassing and validate this proposal.

By supporting the view that the Cenozoic climate evolution is largely controlled by changes in magmatic activity, we put forth a major challenge for future research, because assessing the magmatic forcing on climate requires global-scale, multidisciplinary investigations. Improving data and modeling regarding the strain, topographic and magmatic evolution of plate boundaries, and the redistribution of surface mass across the Earth's surface caused by glacial-interglacial cycles and the associated volcanic response are specifically required to tackle this challenge. Understanding past climate changes will allow us to better assess ongoing modifications and predict future conditions, which in turn should influence policymakers toward responsible, conscious, and shared decisions for adaptation and mitigation options. However, the challenge does not only involve scientists but all individuals that contribute sensitizing the community to the complex network of interactions that control the Earth's climate. This is an essential step toward increasing awareness of the societal impact on our planet.

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