



Model limits on the role of volcanic carbon emissions in regulating glacial–interglacial CO₂ variations

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ABSTRACT

Huybers and Langmuir (2009) proposed that an increase in volcanic activity provoked by ice sheet melting contributed substantially to the deglacial CO₂ increase. Here, their hypothesis is evaluated by prescribing their central, high, and low volcanic CO₂ emission scenarios in the Bern3D carbon cycle-climate model as a perturbation. Reconstructed emissions increase mainly between 15 and 11 ka BP, remain high in the early Holocene and drop after 7 ka BP in all scenarios with total emissions between 181 and 2011 GtC. Simulated increase of atmospheric CO₂ peaks around 6 ka BP at 46 ppm for the central scenario and with a range between 13 and 142 ppm. Modeled carbonate ion concentration in the deep ocean decreases and the calcite saturation horizon shoals on global average by 440 m (150 to 1500 m). Simulated changes in δ¹³C and Δ¹⁴C isotopic signatures are small compared to reconstructed, proxy-based changes over the deglacial period. The comparison of our model results and available proxy evidence suggests a small role for volcanic carbon emissions in regulating glacial–interglacial CO₂ variations, but uncertainties prevent a firm conclusion. A problem with the volcanic emission hypothesis is in the timing of emissions which peak in the early Holocene, a period of decreasing atmospheric CO₂.

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1. Introduction

Many hypotheses (see e.g., Jansen et al. (2007); Köhler et al. (2005)) have been proposed to explain the variations in atmospheric CO₂ concentration of order 100 ppm (parts per million) between glacial and interglacial climate states (Lüthi et al., 2008; Monnin et al., 2004; Petit et al., 1999). Typically, changes in the marine carbon cycle and related ocean–sediment interactions are invoked to explain the CO₂ increase from the Last Glacial Maximum (LGM) to the current warm period. In contrast to this view, Huybers and Langmuir (2009) suggest that an increase in global volcanic activity caused by deglacial ice sheet removal and related changes in pressure could be a major driver for the reconstructed deglacial CO₂ increase.

The goal of this study is to investigate the plausibility of the volcanic CO₂ release hypothesis by Huybers and Langmuir (2009) in the context of available proxy evidence. Specifically, we prescribe their central, low, and high deglacial CO₂ emission scenarios in the Bern3D Earth System Model of Intermediate Complexity and analyze simulated changes in atmospheric CO₂ and its ¹³C and ¹⁴C isotopic signatures as well as the spatio-temporal evolution of carbonate ion concentration, alkalinity, and δ¹³C and Δ¹⁴C of dissolved inorganic carbon in the deep ocean. Results are compared to proxy data.

The conventional hypotheses to explain glacial–interglacial CO₂ variations rely on marine mechanisms and suggest a repartitioning of carbon between the atmosphere, the ocean and ocean sediments, and vegetation and soils. The ocean stores much more carbon than vegetation and soils on land and is by far the largest of the three relatively fast (<1000 years) exchanging carbon reservoirs (atmosphere–ocean–land biosphere). Carbon storage on land is thought to have increased by several hundreds of gigatons of carbon (GtC) from Last Glacial Maximum (LGM) to the current warm period and can thus not explain the LGM to Holocene CO₂ increase. A broad range of proxies, including for example the stable carbon isotope ¹³C signature of atmospheric CO₂ (Elsig et al., 2009; Lourantou et al., 2010), the ¹³C and radiocarbon (¹⁴C) signature of dissolved inorganic carbon (e.g. Duplessy et al., 1988; Oliver et al., 2010; Robinson et al., 2005; Sarnthein et al., 1994), or the carbonate ion concentration in the deep ocean (Yu et al., 2010), indicates large scale changes in ocean circulation, temperature, water mass distribution, and ocean biogeochemistry over glacial–interglacial cycles. However, no consensus has yet emerged on the explanation of the coeval CO₂ changes. The challenge remains to quantify the contribution of identified physical and biogeochemical mechanisms to the CO₂ transient consistent with available proxy information.

Despite the importance of volcanic outgassing of CO₂ on the carbon cycle on geological timescales and early suggestions on their role in regulating glacial–interglacial climate variations (Arrhenius, 1896), changes in global volcanic activity were until recently not considered to explain CO₂ variations on the multi-millennial timescales of glacial–

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interglacial (G/IG) cycles. Several studies addressed changes in local volcanic activity. Tephra-measurements in Antarctica (Narcisi et al., 2010) and references therein) for example reveal past variations in volcanic activity from nearby Antarctic regions. For Iceland, Maclennan et al. (2002) found a pronounced peak in volcanic activity around ~12 kyr ago, exceeding modern rates by a factor of 100. Also for France and Germany, such a relationship could be shown (Nowell et al., 2006). Difficulties for the extrapolation of local records to the globe are the spatio-temporal variability of volcanic activity and, more important, that the quality of the records decreases when going back in time. The available data on volcanic eruptions show a marked observational bias with 80% of the dated eruptions occurring in the last 1000 years (e.g. Fig. 1 in Huybers and Langmuir (2009)). A further step adding additional uncertainty is to estimate CO₂ emissions from estimated volcanic activity. These issues are thoroughly discussed by Huybers and Langmuir (2009).

Huybers and Langmuir (2009) in their stimulating study estimate the global flux-variation of subaerial volcanic carbon over the last 40 kyr by evaluating historical eruption data-sets covering more than 5000 individual volcanic events. Eruption frequencies in regions which are subject to deglacial processes such as ice sheet retreat (especially in the northern hemisphere) are compared to those of regions presumably free of deglacial influences in an attempt to remove observational biases. A considerable increase in volcanic activity of two to six times above background level between 12 kyr and 7 kyr is identified and linearly translated into a corresponding increase in CO₂ flux. Using a box model approach, these authors estimate an increase in atmospheric CO₂ of 60 ppm during the second half of the last deglaciation due to volcanism alone (uncertainty range: 25–130 ppm).

The reason for the strong increase in volcanic activity is thought to have emerged from the retreat of the northern hemisphere ice sheet provoking magma production due to depressurization (Sigmundsson et al., 2010). If the proposed volcanic emission peak during deglaciations is real, volcanism would mediate a positive feedback between increasing atmospheric CO₂, warming, and ice sheet melting. This volcanic deglacial CO₂ scenario is different from the hypotheses that propose a repartitioning of carbon between the atmosphere, ocean, and land biosphere to explain the glacial-to-interglacial CO₂ increase. It involves the net addition of carbon to the atmosphere–ocean–land biosphere system from the earth's upper mantle, a pool that is currently exchanging only little carbon with these other reservoirs.

2. Method

2.1. Model description

Simulations were performed with the Bern3D Earth System Model of Intermediate Complexity (EMIC). The ocean component is a three-dimensional frictional-geostrophic balance ocean model based on Edwards et al. (1998) and further described in Müller et al. (2006). It is run with a horizontal resolution of 36 × 36 grid boxes, with 32 logarithmically spaced layers, and with a time-step of 48⁻¹ yr. Here, the Bering Strait is open and there is a flow of 21 Sv (1 Sv = 10⁶m³s⁻¹) through the Indonesian Passage from the Pacific to the Indian Ocean.

The 2D energy balance model of the atmosphere, described in detail in Ritz et al. (2011), has the same horizontal resolution as the ocean. Zonal uniform diffusivities are applied to simulate horizontal heat and water fluxes. The model distinguishes between shortwave and longwave fluxes between atmosphere–ocean, atmosphere–sea ice and atmosphere–land boundaries. Atmospheric trace gases are considered to be well mixed. The equilibrium climate sensitivity has been tuned toward 3 °C for a doubling of atmospheric CO₂.

The biogeochemical (BGC) component consists of a prognostic representation of the marine carbon cycle, partly following the OCMIP-II protocol (Najjar et al., 1999; Orr and Najjar, 1999), but with prognostic formulations for export production of organic carbon and featuring competition between calcite and opal-producer according to Maier-Reimer

(1993) with Michaelis–Menten formulation of the limiting production terms (Tschumi et al., 2008). A prognostic iron-cycle is included in the model as described in Parekh et al. (2008). In total 14 oceanic tracers are transported.

A 10-layer sediment–diagenesis model (Gehlen et al., 2006; Heinze et al., 1999) is coupled to the ocean as described in detail by Tschumi et al. (2011). It dynamically calculates bioturbation, oxidation, denitrification, dissolution and pore-water diffusion. Modeled solid components which are subject to sediment burial are opal, particulate organic matter (POM), calcite and clay.

We do not model weathering of silicate and carbonate rocks explicitly and assume that weathering of carbonate and silicate rocks remained constant over the past 20,000 years. This assumption appears justified given the large uncertainties and conflicting suggestions on glacial–interglacial changes in weathering (e.g., Munhoven, 2002; Vance et al., 2009) and, more important, by the small potential of weathering to affect atmospheric CO₂ and isotopes on the millennial time scales considered in this study. Technically, the weathering–burial cycle is treated in the following way. Carbon and ¹³C, alkalinity (Alk) and nutrients are lost from the model system by the burial flux of calcium carbonate, opal, and organic matter leaving reactive sediments and entering the lithosphere. This loss is balanced by a corresponding riverine input flux assumed to result from the weathering of silicate and carbonate rocks during model spin-up. The steady-state weathering input flux, diagnosed at the end of the spin-up, is 0.401 GtC yr⁻¹ and has an isotopic ¹³C signature of –8.5% and a carbon to alkalinity molar ratio close to 1. This flux is kept constant throughout the simulation.

A simple 4-box representation of the terrestrial biosphere according to Siegenthaler and Oeschger (1987) is coupled to the atmosphere. A stimulation of net primary productivity (NPP) by elevated CO₂ is parameterized using a logarithmic dependency of NPP on atmospheric CO₂ ($NPP(CO_2) = NPP(CO_{2,0}) \times (1 + \beta \times \ln(CO_2/CO_{2,0}))$). Soil and litter turnover rates, *k*, are taken to vary with global mean surface air temperature, *T* according to $k(T) = k_{10} \times Q_{10}^{(T-10^\circ C)/10^\circ C}$. The response to changing atmospheric CO₂ and temperature has been tuned toward the Lund–Potsdam–Jena dynamic global vegetation model (DGVM) (Sitch et al., 2003). The fertilization factor β is set to 0.3, NPP is 60 GtC yr⁻¹, CO_{2,0} concentration is 278 ppm, and Q₁₀ is 1.3. Carbon isotopes ¹⁴C and ¹³C are represented in all model components (Tschumi et al., 2011).

For the simulation presented here, we use a perturbation approach. The realistic simulation of the transient deglacial CO₂ evolution is beyond the scope of this study. The model is spun up following a similar procedure as described in Tschumi et al. (2011). A relatively well-constrained and documented preindustrial initial state (with 278 ppm atmospheric CO₂) is obtained which is then perturbed with the volcanic forcing. The choice of a preindustrial state as a starting point for our volcanic perturbation experiments seems justified as most of the applied volcanic CO₂ is released during the late termination and the Holocene, a period where atmospheric CO₂ is generally well above ~240 ppm and closer to late Holocene than Last Glacial Maximum concentrations. Using an initial steady state with boundary conditions typically for the Last Glacial Maximum would probably lead to a higher oceanic carbon uptake than the preindustrial steady state as the Revelle factor is lower under lower CO₂ (higher buffering capacity).

The remaining drift in a 40 kyr control simulation is 95 × 10⁻³ ppm kyr⁻¹ for CO₂, 2.0 × 10⁻⁴‰ kyr⁻¹ for δ¹³C and 2.3 × 10⁻²‰ kyr⁻¹ for Δ¹⁴C. All results presented are corrected for this slight drift.

Model performance is discussed for a range of water mass, biogeochemical, and ventilation tracers in earlier publications for the ocean only model (Müller et al., 2006; Parekh et al., 2008; Tschumi et al., 2008, 2011) and for the coupled ocean–EBM model (Ritz et al., 2011). Marine biogeochemical properties and tracer distributions are similar for the ocean only and the coupled model. Here, we present in addition a comparison between the simulated and data-based distribution of carbonate ion concentrations (Fig. 1); concentrations are expressed as deviations from the saturation concentration relative to calcite, ΔCO₃²⁻.

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