

## Glacial-interglacial atmospheric CO<sub>2</sub> variations

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The Vostok ice-core record (Barnola et al., 1987) provided the first evidence of a close relationship between atmospheric CO<sub>2</sub> levels and climate. These observations were confirmed by the later extensions of the records back to 240 kyr BP (Jouzel et al., 1993) and 420 kyr BP (Petit et al., 1999). Both CO<sub>2</sub> and temperature records are represented in Figure 1, together with the CH<sub>4</sub> record.

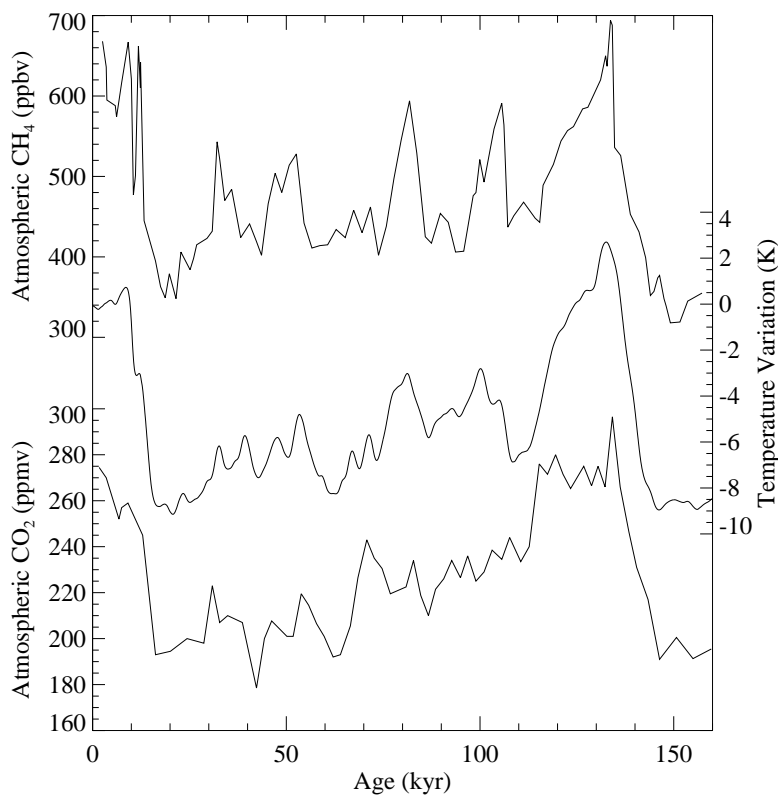


Figure 1: Vostok ice-core records of CH<sub>4</sub> (top – Chappellaz et al., 1990), temperature (middle – Jouzel et al., 1987) and CO<sub>2</sub> (bottom – Barnola et al., 1987) for the past 160,000 years.

Of all the available records on glacial-interglacial changes of the carbon cycle, the atmospheric CO<sub>2</sub> signal represents the strongest constraint, because of its significance at the global scale. There is agreement that the ocean plays a major role in driving atmospheric CO<sub>2</sub> levels at the time scales of interest. The ocean contains about 60 times more carbon than the atmosphere, and the residence time of CO<sub>2</sub> in the atmosphere with respect to air-sea exchange is close to 10 years. On time scales of several tens of years and more, the atmosphere is close to equilibrium with the average surface ocean.

Accordingly, atmospheric CO<sub>2</sub> variations such as those observed here are the result of changes in the surface ocean dissolved inorganic carbon (DIC) or total alkalinity (TA) concentrations: lower CO<sub>2</sub> levels in the atmosphere are associated with increased TA or decreased DIC concentrations in the surface ocean, or with an appropriate combination of the two. Such changes could either be due to global ocean inventory changes, or to changes in the surface-to-deep-sea gradients of the two quantities.

A first group of mechanisms call on a more pronounced surface-to-deep-sea gradient in DIC at the LGM. They mainly focus on the biological pump. A stronger biological pump during glacial times could efficiently reduce atmospheric CO<sub>2</sub>, mainly by shifting large amounts of DIC from the surface to the deep sea. Such a strengthening would result either from a larger nutrient stock at the LGM, or from a better utilisation of the available nutrients at that time. This group includes the *Phosphate Extraction Hypothesis* (Broecker, 1982), the *Denitrification Model* (Broecker, 1982; Keir and Berger, 1983), the *C/P Model* (Broecker, 1982), the *Polar Nutrient Hypothesis* (Knox and McElroy, 1984; Sarmiento and Toggweiler, 1984; Siegenthaler and Wenk, 1984) and the *Iron Fertilisation Hypothesis* (Martin, 1990). Although mainly focusing on DIC changes, these mechanisms also lead to important alkalinity changes, the most important of which are related to perturbations of the carbonate compensation mechanism in the deep sea.

A second group of hypotheses call on alkalinity changes, and mainly focus on the inorganic pump and carbonate compensation in the deep sea. These are the *Rain Ratio Hypothesis* (Broecker, 1982; Berger and Keir, 1984), *variable SiO<sub>2</sub>/CaCO<sub>3</sub> shell ratios* (Dymond and Lyle, 1985), the *Polar Alkalinity Hypothesis* (Broecker and Peng, 1989), the *Coral Reef Hypothesis* (Berger, 1982; Keir and Berger, 1983; Opdyke and Walker, 1992), the *Vertical Nutrient Fractionation Hypothesis* (Boyle, 1988), and *changing weathering rates* (Walker and Opdyke, 1995; Munhoven and François, 1994, 1996)

All of the suggested scenarios have their shortcomings. None of them provides a complete explanation for the observed CO<sub>2</sub> variations. Most of them have consequences that are in contradiction with independent sedimentary evidence.

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