powder size (see Supplementary Information for details). This increase is entirely due to coarsening in the first-step sintering. As nanocrystalline powders in the size range 5–10 nm are becoming increasingly available\(^1,2\), we believe that it should be possible to achieve dense, nanostructured materials of grain size 25–50 nm by exploiting the kinetic ‘window’ that separates grain-boundary diffusion from grain growth. We consider that the simplicity of this approach should also make it useful for detailed explorations of dense nanostructured materials, in order to take advantage of their grain-size-dependent physical properties.

**Methods**

We used a precipitation technique to first obtain Y(OH)\(_3\), which was then calcined at various temperatures from 600 to 800 °C to obtain final Y\(_2\)O\(_3\) powders with a size from 10 to 100 nm. These powders were sifted and pressed into pellets. When doping was desired, Mg and FeO were added and mixed to aqueous slurries of Y\(_2\)O\(_3\) powders; the slurries were gelled by adjusting their pH, then dried and recalcined. The pellets were sintered in a dilatometer or in a box furnace in air. The final density was determined using Archimedes’ method and qualitative microscopy.

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**Supplementary information**

is available on Nature’s World-Wide Web site or as paper copy from the London editorial office of Nature.

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Increases in Antarctic sea ice have been included in three-dimensional ocean model simulations, but their potential to significantly affect atmospheric \( \text{CO}_2 \) in these models may be limited by defects involving excessive low-latitude upwelling of deep water. Toggweiler and Samuels\(^{10} \) have shown from \( ^{14} \text{C} \) comparisons, and Gnanadesikan and Toggweiler\(^{11} \) from silica flux comparisons, that coarse-resolution ocean models all overestimate the amount of deep water upwelling across the main thermocline at low latitudes, in the Indian and Pacific oceans. Previous box models used to explore palaeo-\( \text{CO}_2 \) controls\(^{2-4,14,15} \) also overestimate deep-water upwelling at low latitudes. These models have from 15 to 40 Sv \((1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1})\) of upwelling directly from the deep to the warm surface box, which is much greater than the combined upper limit of ~7 Sv indicated by tracer and transect studies in the Pacific\(^{10} \) and Indian\(^{16} \) oceans.

To investigate the influence of Antarctic sea ice in the absence of low-latitude deep-water upwelling, we employ the biogeochemical ocean model shown in Fig. 1. The fundamental difference between this and other box models of the ocean carbon cycle is that deep waters only return to the surface south of the Antarctic Polar Front (APF), consistent with the "reconfigured conveyor" of Toggweiler and Samuels. We divide the upper ocean into a warm, low-nutrient (APF), consistent with the "reconfigured conveyor" of Toggweiler and Samuels. We divide the upper ocean into a warm, low-nutrient surface (S) box and a cooler, higher-nutrient thermocline (T) box. We also separate the upwelling that feeds the Antarctic Bottom Water (AABW) formation (B) box from that which flows north in the Antarctic surface (A) box to allow for higher nutrient concentrations in AABW. We selected these divisions to optimize the model's representation of the modern distribution of preformed nutrients and export production. However, combining either the B and A boxes or the S and T boxes does not change our results significantly.

Deep waters that upwell along the Antarctic Divergence, mix with surface waters, and then flow north are generally thought to sink again at the APF as part of the low-salinity Antarctic Intermediate Water (AAIW). Although there is considerable uncertainty\(^{19,20}\), it appears that some of this recent deep water remains near the surface north of the APF where it convectively mixes with surface waters of low-latitude origin before sinking. To investigate model sensitivities to uncertainties in the fate of northward-flowing Antarctic surface waters, we allow for surface exposure of an adjustable fraction \((F_s)\) of these waters in the subantarctic (SA) box.

After prescribing temperatures, salinities, water transports, and surface nutrients, and parametrizing the effects of biological production, air–sea gas exchange, and carbonate sediment interaction, we integrate the model to calculate the steady-state distribution of total \( \text{CO}_2 \) (\( \Sigma \text{CO}_2 \)), the \( ^{13} \text{C}/^{12} \text{C} \) ratio (\( \delta ^{13} \text{C} \)), total alkalinity (TA), dissolved \( \text{O}_2 \), and the atmospheric \( \text{CO}_2 \) concentration. Figure 1 depicts a solution to this model using input parameters representing the modern pre-industrial state. To assess the sensitivity of our model to variations in Antarctic ice cover, we have calculated

![Figure 1 Solution from our atmosphere–ocean model for a best-guess modern-preindustrial state. In this state, \( F_s = 0.3 \), where \( F_s \) is the fraction of northward-flowing Antarctic surface waters exposed to the surface in the subantarctic (SA) box. This model consists of an atmosphere and six ocean boxes: a main surface box (S) representing the upper 200 m of water between approximately 50°N and the Southern Subtropical Front (STF), a main thermocline box (T) representing waters between 200 and 1000 m depth with a surface outcrop north of the main surface box, an Antarctic Bottom Water formation box (B) representing the upper 1000 m of water south of the Antarctic Divergence (AD) (~65°S today), an Antarctic surface box (A) representing the upper 250 m of the Southern Ocean south of the APF (~55°S today) and north of the AD, a subantarctic box (SA) representing the upper 500 m of water north of the APF and south of the STF, and a deep box (D) representing the remainder of the world's oceans. Solid straight arrows denote water fluxes and are labelled in Sv. Solid wavy arrows indicate sinking fluxes of organic material and hollow wavy arrows indicate air–sea \( \text{CO}_2 \) fluxes, and both are labelled in \( \text{Gt} \text{ yr}^{-1} \). Numbers above the surface boxes denote \( \text{CO}_2 \) partial pressures in \( \mu \text{atm} \). Values for \( \text{PO}_4 \), \( \Sigma \text{CO}_2 \), TA, dissolved \( \text{O}_2 \), and \( \text{CO}_3^{2-} \) are indicated in \( \mu \text{mol kg}^{-1} \), \( ^{13} \text{C} \) in per mil, temperature \( T \) in °C, salinity \( S \) in p.s.u., and atmospheric \( \text{CO}_2 \) in p.p.m. We assume that 50% of the modern-preindustrial ocean in boxes B and A is covered by sea-ice (represented by jagged lines), leaving a combined area of \( 1.6 \times 10^{13} \text{ m}^2 \) open for the ventilation of upwelled deep waters. We prescribe surface phosphate concentrations in the A, SA, S and T boxes, and the total ocean phosphate concentration, such that the model reproduces the observed deep preformed phosphate (\( \text{PO}_4 \)) concentration of 1.4 \( \mu \text{mol kg}^{-1} \) and the \( \text{PO}_4 \) concentration of intermediate waters penetrating the deep ocean, also equal to 1.4 \( \mu \text{mol kg}^{-1} \) (ref. 30). The model consumes excess phosphate in these four surface boxes at revised Redfield proportions of ~175 \( \text{O}_2 :1 \) 127 C :1 16 N :1 P. As the resulting organic matter sinks, 80% of the main surface flux is oxidized in the thermoline box and the remainder of all three fluxes in the deep box. We assume inorganic to organic carbon production ratios of 0, 1:20, 1:10 and 1:3 in the A, SA, S and T boxes respectively. The model remineralizes all sinking calcium carbonate in the deep box. We calculate fluxes across the air–sea interface assuming a perfectly mixed atmosphere and using \( \text{CO}_2 \) invasion rates of 0.15 (B, A, SA, T) and 0.05 \( \text{S m} \text{ m}^{-2} \text{ yr}^{-1} \) \( \text{CO}_2 \text{ atm}^{-1} \), and \( \text{O}_2 \) gas transfer velocities of 45 (B, A, SA, T) and 15 (s cm) \text{ h}^{-1}. We maintain a constant \( \text{CO}_2 \) concentration in the deep box by dissolving or precipitating an appropriate amount of CaCO$_3$ (ref. 8). We have adjusted the mixing between the thermoline and surface boxes to produce a reasonable partitioning of biological productivity between these boxes.

![Diagram showing the model's representation of atmospheric and oceanic carbon distributions.](https://example.com/diagram.png)
steady-state solutions using these parameters and varying the ice-free surface area south of the APF from $1.6 \times 10^{12}$ m$^2$ down to zero. Figure 2a shows the atmospheric CO$_2$ concentration for these solutions for four cases with different values of $F_a$. The decrease in atmospheric CO$_2$ of 67 parts per million (p.p.m.) for the best-guess conditions shown in Fig. 2 is associated with a 92% decrease in the sea-to-air CO$_2$ flux south of the APF, and a 1.8% increase in the deep $\Sigma$CO$_2$ concentration, with no change in the nutrient distribution. The deep $\Sigma$CO$_2$ increase contributes to the dissolution of $2.5 \times 10^{16}$ mol of CaCO$_3$, which produces a 1.6% increase in whole ocean alkalinity and contributes 13 p.p.m. of the total atmospheric CO$_2$ effect. Although not enough to produce large changes in glacial lysolynci depths, this alkalinity–ice relationship is consistent with observations of a global carbonate-sediment preservation spike during deglaciation.

In contrast to atmospheric CO$_2$, the modelled deep O$_2$ concentration is not sensitive to ice coverage except when the outcrop area becomes very small (Fig. 2b). The change in deep O$_2$ from a modern-preindustrial estimate of 191 $\mu$mol kg$^{-1}$ to a possible glacial value of 123 $\mu$mol kg$^{-1}$ (Fig. 2b) would not produce deep anoxia. Our model also supports earlier suggestions$^{2,27}$ that sea ice is an important influence on glacial Antarctic surface $\delta^{13}$C values. Fractionation effects associated with both net and gross air–sea CO$_2$ fluxes work to make the Antarctic surface box richer in $\delta^{13}$C values. As these fluxes decrease with increasing ice cover, the Antarctic surface $\delta^{13}$C value decreases by 0.7% (Fig. 2b). This decrease is similar to that inferred from measurements on planktonic foraminifera$^{2,28}$, and is in contrast to the increases in Antarctic surface $\delta^{13}$C implied by models that invoke increases in high-latitude nutrient utilization or decreases in high-latitude vertical mixing.

Atmospheric $\delta^{13}$C increases with increasing sea ice by 0.9‰ in the best-case model scenario. If we include an input of 500 gigatons of terrestrial carbon$^{23}$, and reduce surface temperatures by a maximum of 5 $^\circ$C (ref. 24) and less in boxes that are already near the freezing point—the atmospheric $\delta^{13}$C value returns to within 0.1‰ of the modern-preindustrial value. Measurements$^{25}$ indicate a sharp dip in atmospheric $\delta^{13}$C of $\sim$ 0.5‰ at the start of termination I, followed by a more gradual increase to a pre-industrial value several tenths of a per mil above that at the Last Glacial Maximum (LGM). Although our model does not reproduce this overall shift, it does suggest that the temporal behaviour of atmospheric $\delta^{13}$C during deglaciation could be explained by an initial meltback of Antarctic sea ice followed by a more gradual growth in terrestrial biomass and increase in surface temperatures.

Although it is difficult to estimate the extent of glacial Antarctic sea ice, suggestions from sediment proxies of significantly increased coverage justify the investigation of its potential implications for atmospheric CO$_2$. There is solid evidence from ice-rafted volcanic detritus$^{26}$, and support from fossil plankton assemblages$^{27}$, that wintertime Antarctic sea ice extended to or beyond the modern APF during the LGM. However, summertime LGM ice limits are not as well constrained. Early estimates of significantly increased coverage based on sediment types$^{26}$ have been countered by recent planktonic analyses that suggest summertime Antarctic sea ice at the LGM may not have been much more extensive than today$^{27}$. For our purposes, we expect the winter ice extent to have the main effect on deep-water ventilation, as stratification and biological productivity during summer independently limit the outgassing of CO$_2$ and uptake of O$_2$. François et al.$^{24}$ used sediment $\delta^{15}$N and opal data to estimate that increased stratification south of the APF during glacial times led to a 70% utilization of nutrients. This would be sufficient to remove 196 $\mu$mol C kg$^{-1}$ ($0.7 \times 127 \times 2.2 \mu$mol phosphate per kg) from the Antarctic surface waters, and thus prevent summertime CO$_2$ outgassing in our model. Furthermore, the extreme reduction in vertical mixing implied by the tenfold decrease in nutrient inputs estimated by François et al.$^{24}$ suggests that a large amount of ice remained in this region throughout the summer.

The results shown in Fig. 2 should be viewed as an indication of a potentially important mechanism in the ocean–atmosphere carbon system, rather than as an absolute prediction of the magnitude of the Antarctic sea-ice effect. However, we note that its magnitude is fairly robust with respect to variations in the assumed parameters. In addition to the sensitivities to ice-free area and $F_a$ shown in Fig. 2a, the best-guess CO$_2$ difference changes by only $\sim$ 9 p.p.m. or $\pm$ 10 p.p.m. if the entrainment of low-latitude waters in AAIW is respectively doubled or reduced to zero. Although our model illustrates the behaviour of the ocean in the limit of no low-latitude deep-water upwelling, it is likely that there is some finite amount of diapycnal flow through the main thermocline in the real ocean. However, if we include a high-latitude sinking, low-latitude upwelling term of 10 Sv in our model, the total atmospheric CO$_2$ difference decreases by only 8 p.p.m. to a total of 59 p.p.m.

In addition, we have not tried to simulate increases in nutrient utilization in the subantarctic, yet in our model such changes could significantly affect the amount of CO$_2$ entering the deep ocean. If we reduce subantarctic surface nutrients after increasing Antarctic sea ice to generate an increase from 1.5 to 2.8 mol C m$^{-2}$ yr$^{-1}$ in subantarctic export production (slightly less than that proposed by François et al.$^{25}$), the CO$_2$ drawdown increases by 15 p.p.m. to a total of 82 p.p.m. A final perturbation to consider is that of temperature-driven solubility changes. After reducing surface temperatures as described above, our model predicts an additional CO$_2$ drawdown of 27 p.p.m., corresponding to a Harvardton Bear Index$^{26}$ of 0.2. The combined effects of ice cover, subantarctic productivity, and temperature simulated by our model are

Figure 2 Steady-state model solutions for different ice coverages south of the Antarctic Polar Front (APF). We vary the exposed sea surface in the B and A boxes such that the same fraction of total surface is ice-free in each. The x-axis values represent the sum of exposed area in these two boxes, and are scaled by their square root to expand the left side of the plot. We use $1.6 \times 10^{12}$ m$^2$ as a modern estimate of the ice-free area south of the APF, and $1.6 \times 10^{13}$ m$^2$ to illustrate a possible glacial value. a, Atmospheric CO$_2$ using modern-preindustrial parameters and different fractions ($F_a$) of Antarctic surface waters entering the subantarctic box. Based on the evidence of Molinelli$^{19}$ showing the importance of subsurface transport of Antarctic waters relative to vertical mixing as the subantarctic source of AAIW, we have chosen a relatively low $F_a$ value of 0.3 to use as a best-guess case. b, Atmospheric CO$_2$, deep O$_2$, and Antarctic surface $\delta^{13}$C using $F_a = 0.3$. 

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sufficiently large to ensure that an 80 p.p.m. decrease in glacial atmospheric CO₂ can still be attained even after including the counteracting effects of terrestrial biomass changes and salinity-driven solubility effects3. Toggweiler recently developed a model that similarly invokes a reduction in deep-water ventilation to explain the low glacial CO₂ levels. Whereas he generates reduced ventilation by decreasing the vertical exchange between deep and Antarctic surface waters, we suggest that reduced ventilation was driven by limitations to air–sea gas exchange imposed by increased sea ice in this region.

**Interferometric radar measurements of water level changes on the Amazon floodplain**


Measurements of water levels in the main channels of rivers, upland tributaries and floodplain lakes are necessary for understanding flooding hazards, methane production, sediment transport and nutrient exchange. But most remote river basins have only a few gauging stations and these tend to be restricted to larger river channels. Although radar remote sensing techniques using interferometric phase measurements have the potential to greatly improve spatial sampling, the phase is temporally incoherent over open water and has therefore not been used to determine water levels. Here we use interferometric synthetic aperture radar (SAR) data1–3, acquired over the central Amazon by the Space Shuttle imaging radar mission4, to measure subtle water level changes in an area of flood vegetation on the Amazon floodplain. The technique makes use of the fact that flooded forests and floodplain lakes with emergent shrubs permit radar double-bounce returns from water and vegetation surfaces5,6, thus allowing coherence to be maintained. Our interferometric phase observations show decreases in water levels of 7–11 cm per day for tributaries and lakes within ~20 km of a main channel and 2–5 cm per day at distances of ~80 km. Proximal floodplain observations are in close agreement with main-channel gauge records, indicating a rapid response of the flood plain to decreases in river stage. With additional data from future satellite missions, the technique described here should provide direct observations important for understanding flood dynamics and hydrological exchange between rivers and flood plains.

Climatically driven, seasonal changes in river water levels (river stages) govern a wide range of hydrologic, geomorphological and ecological processes. Hydrologic modelling of the Amazon flood wave predicts discharge only along the main channel (the main stem), and suggests that up to 30% of mainstem flow exchanges with the floodplain7. On the basis of transport models, the annual sediment exchange between the main channel and floodplain is more than twice the flux through the most downstream river gauge at Obidos8. However, the models do not describe the sources or residence times of the floodplain water; these are key variables for understanding biological productivity and sedimentation. For example, observations of a small local catchment demonstrate that early in the water year, the lake contains nearly 70% river water whereas later in the water year, local runoff and other sources increase lake stage, preventing rising-flood-stage river water from entering and exchanging with the lake9. Because very few of the ~8,000 Amazon floodplain lakes are gauged, the generality of these observations is unclear10.

Remotely sensed observations of the water surface provide an alternative to permanent gauging. Satellite radar altimetry promises a stage elevation accuracy of about 10 cm: but because altimetry is a profiling and not an imaging technique, it is applicable only to water