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A Paleo-Perspective on Changes in Atmospheric CO_2 and Climate

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The goal of the United Nations Framework Convention on Climate Change is to stabilize atmospheric CO_2 and other greenhouse gases (GHGs) to prevent "dangerous" anthropogenic interference with climate. CO_2 is the most important of the anthropogenic GHGs in terms of radiative forcing, and its stabilization presents major challenges to society. To stabilize atmospheric CO_2 concentration at any given level, it is necessary to define trajectories of allowable CO_2 emissions. This calculation requires a thorough understanding of the carbon cycle—that is, the various terrestrial and oceanic processes that control the amount of CO_2 that stays in the atmosphere.

Timescales of carbon cycle processes range from months to centuries for carbon exchange between the atmosphere, land biosphere, and ocean to millennia for ocean-sediment interactions. Processes that operate on timescales longer than a few years cannot be investigated by experimental field studies, and long-term instrumental observations are missing for many variables. Paleodata (especially measurements on ice cores, marine and lake sediments, corals, tree rings, and historical documents) provide unique information on the magnitude of natural CO_2 variability, the processes regulating atmospheric CO_2 and their associated timescales (including multimillennial timescales), and the magnitude of carbon cycle–climate feedbacks.

The longest available ice-core records (from Vostok, Antarctica [Petit et al. 1999] and Dome Fuji, Antarctica [Kawamura et al. 2003]) demonstrate that atmospheric CO_2 concentration today is higher than at any time during (at least) the past 420,000 years.

The search for mechanisms driving the observed glacial-interglacial changes has led to the identification of a range of processes that contribute to the control of atmospheric CO_2 concentration and climate. The detection of abrupt, decadal-scale climatic changes influencing large regions has fueled the concern that anthropogenic GHG emissions may trigger future abrupt climate changes. Changes in CO_2 and climate during the Holocene, and during the last millennium of the Holocene, are less spectacu-

lar but provide important information on the linkages between the carbon cycle and climate.

Glacial-Interglacial Variations

The Vostok and Dome Fuji ice-core records reveal that atmospheric CO_2 variations over the past four glacial cycles were confined to the range between ~ 180 parts per million (ppm) (around glacial maxima) and ~ 280 ppm (typical for interglacial periods) (Petit et al. 1999; Kawamura et al. 2003). Thus, preindustrial atmospheric CO_2 concentrations were consistently lower than today's value of 370 ppm and dramatically lower than the projected concentration range at year 2100 (450 to 1,100 ppm) reported by the Intergovernmental Panel on Climate Change (IPCC) (Joos et al. 2001; Prentice et al. 2001).

A success of the greenhouse theory, first established in the 19th century (Arrhenius 1896) and of today's climate models is that both the global warming over the industrial period (Houghton et al. 2001) and the widespread cold conditions of the last glacial maximum (LGM) (Ganopolski et al. 1998; Weaver et al. 1998; Kitoh et al. 2001; Kim et al. 2002; Hewitt et al. 2003; Shin et al. 2003) can be consistently explained by the radiative forcing due to changes in atmospheric GHG content and other factors. Detailed comparison of temperature proxies and CO_2 during the last glacial-interglacial transition, however, suggests that Antarctic temperature started to rise before atmospheric CO_2 (Figure 7.1). This finding is consistent with the view that natural CO_2 variations constitute a feedback in the glacial-interglacial cycle rather than a primary cause (Shackleton 2000). Changes in the Earth's orbit around the Sun are the pacemaker for glacial-interglacial cycles (Hays et al. 1976; Berger 1978), but these rather subtle orbital changes must be amplified by climate feedbacks in order to explain the large differences in global temperature and ice volume, and the relative abruptness of the transitions between glacial and interglacial periods (Berger et al. 1998; Clark et al. 1999).

Biogeochemical cycles play an important role for the amplification of orbital changes. Model simulations suggest that the direct radiative forcing by atmospheric CO_2 and CH_4 concentrations may have contributed up to half of the observed glacial-interglacial surface temperature difference at a global scale (Broccoli and Manabe 1987; Gallée et al. 1992; Shin et al. 2003). Other major factors involved in maintaining cold conditions during glacial periods include the water vapor feedback, the high albedo of the continental ice sheets (Broccoli and Manabe 1987; Hewitt and Mitchell 1997), the high albedo (especially when snow covered) of extensive nonforested regions at high latitudes (Gallée et al. 1992; Levis et al. 1999; Yoshimori et al. 2001; Wyputta and McAvaney 2002), and the reflection of shortwave radiation by the greatly enhanced atmospheric content of mineral dust (Claquin et al. 2003)—itself a consequence of reduced vegetation cover (Mahowald et al. 1999; Werner et al. 2002). It is plausible that such biogeochemical and biophysical feedbacks will amplify the direct anthropogenic greenhouse gas forcing, just as they have amplified orbital changes in the past.

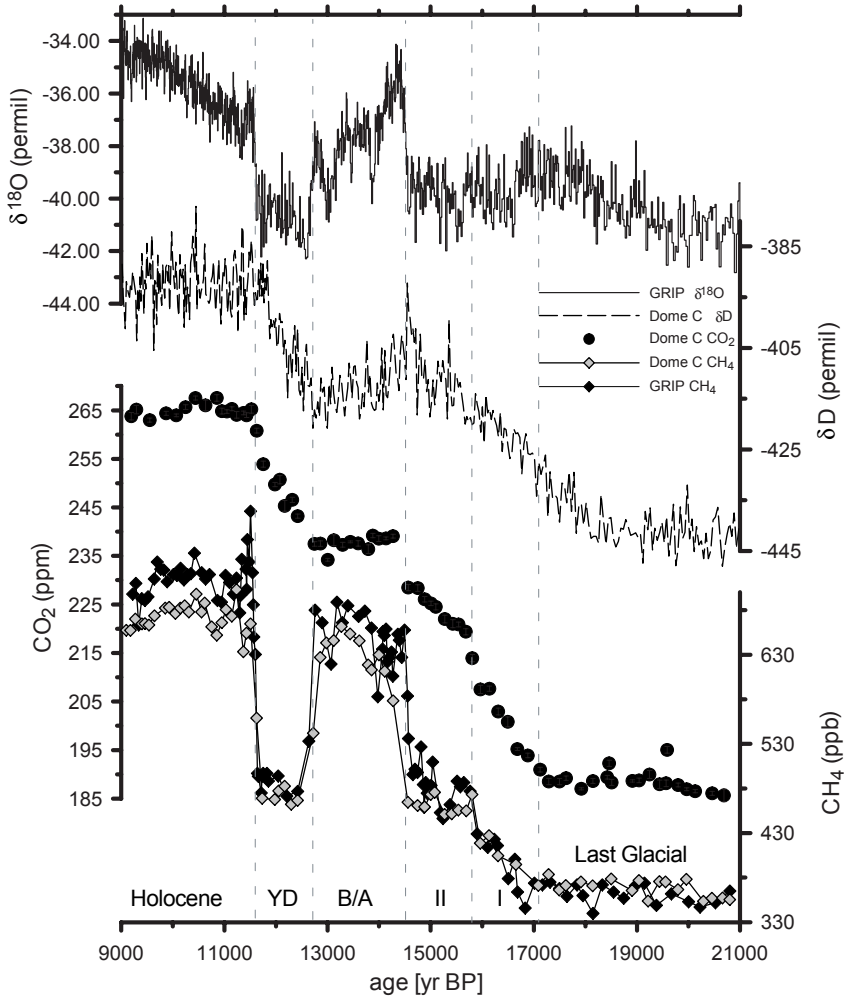

Figure 7.1. The evolution of proxies for local temperature in Greenland ($\delta^{18}O$, top line) and Antarctica (δD, dash) and the atmospheric concentration of the two greenhouse gases CO_2 (dots) and CH_4 (diamonds) over the last glacial-interglacial transition (Monnin et al. 2001). The records of Antarctic temperature and atmospheric CO_2 are highly correlated and can be divided into four phases as indicated by dashed vertical lines. The rapid (decadal-scale) variations in CH_4 at the beginning of the Bølling/Allerød (B/A) and at the beginning and end of the Younger Dryas (YD) are coeval with abrupt temperature changes in the North Atlantic region. Temperature changes in Antarctica are smaller and less abrupt than in Greenland, and temperature changes in Greenland and Antarctica are asynchronous. Measurements are from the EPICA ice core drilled at Dome Concordia, Antarctica, and from the GRIP ice core drilled at Summit, Greenland. The timescales of both cores have been synchronized to the GRIP scale.

The Earth's orbit around the Sun can be calculated with high precision for the future as well as the past (Berger 1978; Berger and Loutre 1991). Future climate changes can therefore be forecast on a multimillenial time scale with some confidence. Even under a natural CO_2 regime (i.e., with the global temperature-CO_2 correlation continuing as in the Vostok ice core), the next glacial period would not be expected to start within the next 50,000 years (Loutre and Berger 2000; Berger and Loutre 2002). The Holocene will thus last longer than the previous (Eemian) interglacial, which endured for only about 10,000 years. This difference is because orbital eccentricity was much higher in the Eemian than today. With low orbital eccentricity, the effects of precession are minimized, and extreme cold-northern-summer orbital configurations like that of the last glacial initiation at 115 kilo years before present (ka BP) do not occur. Sustained high atmospheric greenhouse concentrations, comparable to a mid-range CO_2 stabilization scenario, may lead to a complete melting of the Greenland ice cap (Church et al. 2001) and further delay the onset of the next glacial period (Loutre and Berger 2000).

The mechanistic explanation for the observed glacial-interglacial CO_2 variations remains a difficult attribution problem (Broecker and Henderson 1998; Archer et al. 2000). Benthic carbon-isotope evidence from marine sediments, and terrestrial carbon accounting based on pollen data from terrestrial sediments, suggest that terrestrial carbon storage was ~ 300–700 PgC less at the LGM than during the Holocene (Shackleton 1977; Curry et al. 1988; Duplessy et al. 1988; Van Campo et al. 1993; Bird et al. 1994; Crowley 1995; Peng et al. 1998; Beerling 1999), implying that the extra stored carbon during glacial periods must be in the ocean and not on land. Numerous oceanic processes have been identified that could have contributed to the low glacial concentrations of CO_2 (Archer et al. 2000; Sigman and Boyle 2000; Stephens and Keeling 2000). One explanation conceived in the 1980s (Knox and McElroy 1984; Sarmiento and Toggweiler 1984; Siegenthaler and Wenk 1984), and recently supported by nitrogen-isotope data (Francois et al. 1998; Crosta and Shemesh 2002), invokes a more efficient utilization of macronutrients in the Southern Ocean during glacial times, leading to higher rates of carbon export from the surface and thus to increased carbon storage at depth, reducing the equilibrium concentration of CO_2 at the ocean surface. This scenario could have been realized by an increased input of iron (which is a limiting micronutrient for phytoplankton, especially diatom, growth in the Southern Ocean today [Martin et al. 1994; Boyd et al. 2000]) provided by an enhanced supply and transport of mineral dust (Andersen et al. 1998; Mahowald et al. 1999), possibly in concert with a slower rate of surface-to-deep mixing in the Southern Ocean, in glacial time. This mechanism may account for up to half of the glacial-interglacial difference in atmospheric CO_2 concentration (Levèvre and Watson 1999; Watson et al. 2000; Bopp et al. 2003). It is unlikely that the iron hypothesis could account for more than half of the difference. The Antarctic ice-core records show that about half of the CO_2 lowering during the last glacial period took place before high dust fluxes appeared in the Antarctic (Petit et al. 1999; Röthlis-

berger et al. 2002). Similarly, substantial changes in CO_2 occurred during the LGM-Holocene transition after the high fluxes disappeared (Röthlisberger et al. 2002). The initial slow CO_2 decline during the first part of the last glacial requires a different explanation, possibly involving ocean-sediment interaction with a timescale of millennia (Archer et al. 1999). The subsequent lowering of atmospheric CO_2 by a further ~ 30 ppm occurred more rapidly, coeval with the increase of Antarctic dust.

Recent analysis of dust-storm data has shown that the natural component of the contemporary dust loading in the atmosphere does exceed 75 percent (Tegen et al. 2003), although a substantial fraction of the contemporary dust was formerly attributed to human activities such as overgrazing and construction (Tegen and Fung 1995). Future dust supply to the Southern Ocean and other high-nutrient low-chlorophyll regions could change through changes in vegetation cover and surface characteristics in response to changes in climate, atmospheric CO_2, and land use activities or through changes in atmospheric transport. The resulting changes in marine iron supply could provide a small feedback to increasing atmospheric CO_2.

Abrupt Climatic Changes

Greenland ice-core records show abrupt, decadal-scale temperature changes locally of up to 16 K amplitude (Schwander et al. 1997; Severinghaus et al. 1998; Lang et al. 1999; Severinghaus and Brook 1999) during the last glacial period and the transition to the present interglacial. These are known as the Dansgard-Oeschger (D-O) events (Dansgaard et al. 1982; Oeschger et al. 1984; Broecker et al. 1985; Clark et al. 2002) (Figure 7.2). Isotopic sedimentary and pollen records from lakes and marine sediments (Eicher et al. 1981; Ruddiman and McIntyre 1981; Bond et al. 1993; Yu and Eicher 1998; Ammann et al. 2000; Sánchez Goñi et al. 2000; Baker et al. 2001; Prokopenko et al. 2001; Sánchez Goñi et al. 2002; Tzedakis et al. 2002) and European and Asian loess records (Ding et al. 1999; Ye et al. 2000; Porter 2001; Rousseau et al. 2002) demonstrate that substantial effects of these abrupt climate changes extended over the North Atlantic region and beyond.

Concomitant temperature changes recorded in Antarctica are smaller, less abrupt, and asynchronous to the Northern Hemisphere changes (Indermühle et al. 2000). This asynchrony has been explained (Stocker and Johnsen 2003) by a reduction in the North Atlantic Deep Water formation rate and oceanic heat transport into the North Atlantic region (Stocker 2000), a phenomenon that, in combination with the large heat capacity of the Southern Ocean, produces cooling in the North Atlantic and warming in the Southern Hemisphere (Mikolajewicz 1996; Marchal et al. 1999a, b).

The lowest temperatures in Greenland are associated with the surging of large amounts of ice, recorded as "Heinrich events," which are characterized by the widespread appearance of ice-rafted debris in North Atlantic sediments (Bond et al. 1993; Bond and Lotti 1995). The ice release provides a plausible mechanism (Alley et al. 1999)

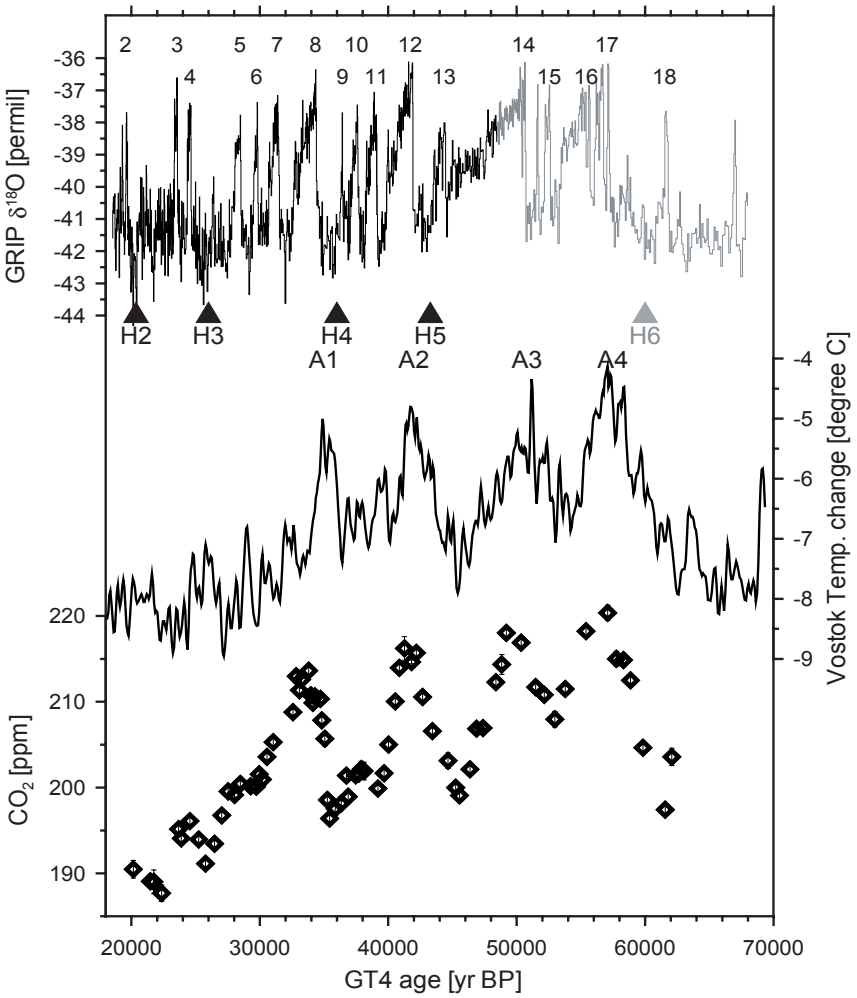

Figure 7.2. The evolution of atmospheric CO_2 concentration and Greenland and Antarctic temperature as indicated by $\delta^{18}O$, during the period from 70 ka BP to 20 ka BP (Indermühle et al. 2000). $\delta^{18}O$ values are from ice drilled at Summit during the Greenland Ice Core Project (GRIP) and Vostok, Antarctica, and CO_2 data are from Taylor Dome, Antarctica. All data are plotted on the GT4 chronology of Vostok. Dansgaard/Oeschger events 2 to 18 (warm interstadials in the Greenland $\delta^{18}O$ record), the Heinrich events H2 to H6 (filled triangles), recorded as ice-rafted debris in marine sediments in the North Atlantic, and the Antarctic warm periods A1 to A4 are indicated. The GRIP, Vostok, and Taylor Dome ice cores were synchronized based on methane measurements. The location of Heinrich events is based on the synchronization of the GRIP ice core to North Atlantic deep sea cores (Bond and Lotti 1995). The timescale is tentative for the first part of the record indicated by light gray in the $\delta^{18}O$ curve.

for a collapse of the thermohaline circulation (THC) as the meltwater input into the North Atlantic stabilizes the water column. Benthic carbon-isotope (Sarnthein et al. 1994) suggests that the deep Atlantic Ocean during the cold periods was filled with nutrient-rich water of Southern Hemisphere origin, and the atmospheric records of ^{14}C and ^{10}Be suggest that the ocean's ventilation was slowed (Hughen et al. 2000; Muscheler et al. 2000; Marchal et al. 2001).

A plausible explanation for the observed changes is the following sequence (Schmittner et al. 2002): Ice accumulates during the interstadial warm phases on the Laurentide ice sheet; this leads to an instability, triggering a massive iceberg discharge into the North Atlantic; the freshwater input causes a temporary breakdown of the THC, with cooling in the north and warming in the south; finally the THC resumes, perhaps in response to slow climatic changes in the Southern Ocean (Knorr and Lohmann 2003), when the freshwater perturbation has dissipated. This mechanism has been questioned by a high-resolution ice-core record that shows that the Antarctic cold reversal may have started earlier than the Bølling/Allerød (B/A) warming (Morgan et al. 2002), implying that the Antarctic cooling was not caused by the restarting of the THC in the Northern Hemisphere and thus suggesting a more important role for the tropics and Southern Hemisphere in controlling abrupt climate changes. A precise determination of hemispheric temperature relationships from the isotopic signatures in ice, however, remains challenging, as factors other than temperature, such as changes in the seasonality of precipitation, can affect the isotopic composition of the ice, and noise in the records may bias results.

Most current climate models, when forced with scenarios of increasing GHG concentrations, show a decrease in the THC and a reduction in the poleward heat transport by the North Atlantic (Cubasch et al. 2001). Idealized model simulations also suggest that the THC is subject to instabilities (as the palaeodata confirm) and that the North Atlantic Deep Water formation might collapse in a warming world (Bryan 1986; Stocker and Schmittner 1997) and remain in a collapsed state, owing to hysteresis, for many centuries. Unfortunately, the existence of critical thresholds and nonlinearities seriously limits our ability to predict the behavior of the THC (Knutti and Stocker 2002). Long-term ocean monitoring may be necessary if we are to reliably detect changes in the THC. Tracers related to the carbon cycle, such as dissolved O_2, are sensitive to circulation changes. Available observations suggest a decreasing inventory of dissolved O_2 in the ocean comparable to that predicted by modeling the effects of global warming on the ocean circulation (Plattner et al. 2001, 2002; Bopp et al. 2002). This effect is also present as an increase in atmospheric O_2 concentration, requiring a correction to the land-atmosphere partitioning of anthropogenic CO_2 (Bopp et al. 2002; Keeling and Garcia 2002; Plattner et al. 2002), estimated from observed atmospheric CO_2 and O_2 trends (Keeling and Shertz 1992; Prentice et al. 2001). An extended monitoring of dissolved O_2 has been proposed to detect changes in ocean circulation and to improve estimates of the carbon sinks (Joos et al. 2003b).

Despite the large fluctuations in climate, atmospheric CO_2 varied <20 ppm during the large Dansgaard/Oeschger events associated with Heinrich events and less than 10 ppm during the smaller Dansgaard/Oeschger events (Figure 7.2) (Stauffer et al. 1998; Indermühle et al. 2000). CO_2 variations were ~ 30 ppm around the Younger Dryas (YD) cold interval (Figure 7.1), during which the North Atlantic THC was greatly reduced (Sarnthein et al. 1994). Concomitant changes in methane and in N_2O were about 100–200 ppb and ~30 ppb, respectively. Ocean model simulations reproduce the observed behavior of the coupled carbon-climate system during abrupt events of this type. Thus, when the North Atlantic THC is forced to collapse by imposing a meltwater pulse in an ocean model, the simulated consequences include a small (10 ppm) temporary increase in atmospheric CO_2, strong cooling in the North Atlantic region, a slight warming in the Southern Hemisphere, an increase in nutrient-rich water in the Atlantic, higher $^{14}C/^{12}C$ ratios in the atmosphere (Marchal et al. 1999a, b; Delaygue et al. 2003), and a 10 parts per billion (ppb) reduction in atmospheric N_2O (Goldstein et al. 2003), that is, about a third of the observed N_2O changes (Flückiger et al. 1999). Similarly, a relatively small positive feedback between atmospheric CO_2 and ocean circulation changes is found in global warming simulations in which the rate of North Atlantic Deep Water formation is reduced or even collapsed (Joos et al. 1999; Plattner et al. 2001). Thus, paleodata and model simulations agree that possible future changes in the North Atlantic Deep Water formation rate would have only modest effects on atmospheric CO_2. This finding does not, however, preclude the possibility that circulation changes in other ocean regions, in particular in the Southern Ocean, could have a larger impact on atmospheric CO_2 (Greenblatt and Sarmiento, Chapter 13, this volume).

This analysis does not consider the possible contribution of vegetation dieback to atmospheric CO_2. Vegetation dieback may have contributed to the observed changes in atmospheric CO_2 concentration during the YD and the Dansgaard/Oeschger events (Scholze et al. 2003) and could have larger effects if an abrupt collapse of the North Atlantic THC occurred during a warm-climate regime.

The Holocene

Atmospheric CO_2 concentration varied slowly during the preindustrial Holocene (Figure 7.3). Between 11 and 8 ka, atmospheric CO_2 fell gradually from 265 to 260 ppm, then it increased again toward the preindustrial level of 280 ppm. The climate event at 8,200 yr BP, comparable to a small Dansgaard/Oeschger event and associated with a temperature drop of 1.5 to 4 K in the North Atlantic region and 4 to 8 K in central Greenland (Barber et al. 1999), left no trace in the available low-resolution CO_2 records (Figure 7.3). A preliminary set of twelve $\delta^{13}C$ measurements was used by Indermühle et al. (1999) to estimate possible causes of the observed long-term variations using a double deconvolution technique (Joos and Bruno 1998). This analysis suggested that the observed variations were caused mainly by uptake and subsequent release of ter-

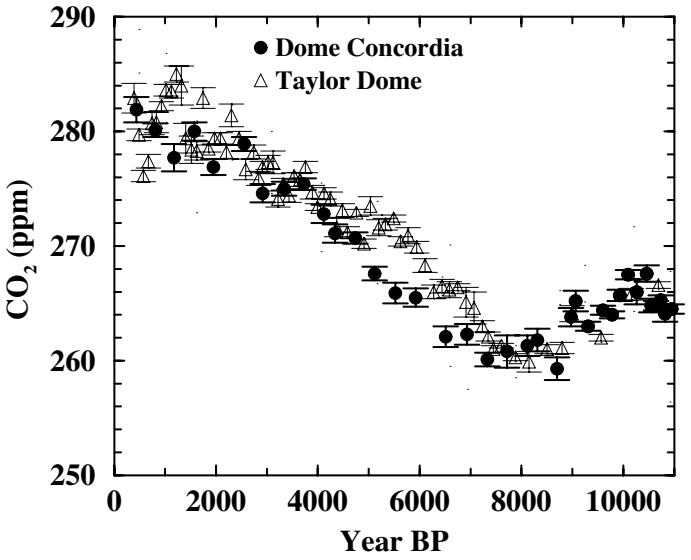

Figure 7.3. Holocene variations in atmospheric CO_2 concentration, from measurements on air entrapped in ice from Taylor Dome, Antarctica (Indermühle et al. 1999) and from Dome Concordia (Flückiger et al. 2002). The difference between the two cores, notably between 4 and 8 kyr, is most likely due to uncertainties in the age scales of the two cores (Stauffer et al. 2002).

restrial organic carbon (Indermühle et al. 1999). An alternative explanation (Broecker et al. 2001; Broecker and Clark 2003) is that the steady CO_2 rise after 8 ka is due to the $CaCO_3$ compensation mechanism, responding to the earlier extraction of carbon from the atmosphere by terrestrial uptake during the period after the LGM. This mechanism depends on ocean-sediment interactions and has an appropriately long time constants of 5 ka and longer (Archer et al. 1999). Unfortunately, uncertainties in the existing ice core $\delta^{13}C$ (Indermühle et al. 1999) and marine sediment data (Broecker and Clark 2003) do not allow us to discriminate reliably between the two hypotheses based on observations alone.

Other information sources provide circumstantial evidence. The benthic carbon isotope record (Curry et al. 1988; Duplessy et al. 1988) constrains the terrestrial uptake during the last glacial-interglacial transition and, hence, constrains the contribution by the CaCO3 compensation mechanism to within 4 to 10 ppm (Joos et al. 2003a). Simulations with land biosphere models (François et al. 1999; Brovkin et al. 2002; Joos et al. 2003a; Kaplan et al. 2002) and pollen-based estimates (Adams and Faure 1998) yield ambiguous results ranging from a terrestrial release of 90 PgC to an uptake of 130 PgC during the late Holocene.

The Lund-Potsdam-Jena (LPJ) dynamic global vegetation model (Sitch et al. 2003) coupled to the atmosphere-ocean-sediment carbon cycle component of the Bern CC model has been used to compare the conflicting explanations for the Holocene CO_2 rise (Joos et al. 2003a). The model was forced with a 21 ka-long time series of "snapshot" simulations from the Hadley Center climate model (Kaplan et al. 2002) or the NCAR climate model and reconstructed atmospheric CO_2 (Monnin et al. 2001) during the transition until 11 ka BP. Afterward, atmospheric CO_2 was simulated. The entire Holocene ice-core CO_2 record is matched within a few ppm. A sensitivity analysis in combination with available observations suggests that $CaCO_3$ compensation, sea surface warming, and a terrestrial release, in response to land use and desertification, contributed about equally to the observed 20 ppm rise after 8 ka BP. The modeled CO_2 decrease during the early Holocene is, in agreement with earlier suggestions, mainly due to the establishment of boreal forest in formerly glaciated areas, partly compensated for by $CaCO_3$ compensation. Simulated LGM-Holocene terrestrial uptake is 800 PgC, slightly higher than the 300 to 700 PgC range derived from the benthic isotope record. These simulations took into account the dynamic land area changes caused by sea-level rise and ice retreat. Carbon sequestration by regrowth and soil establishment on formerly ice-covered areas is almost compensated for by carbon loss in response to sea-level rise and climate change on areas that were ice-free at the LGM. Increasing carbon storage on land in these simulations was principally a consequence of climate change (forced by orbital variations, CO_2, and ice extent), in combination with direct physiological effects of rising CO_2 concentration on plant productivity and water-use efficiency. These direct CO_2 effects were responsible for more than 80 percent of the simulated difference in global carbon storage between the Holocene and the LGM.

Existing uncertainties in the ice-core $\delta^{13}C$ and marine sediment records need to be reduced to further constrain the quantitative contributions of individual mechanisms to the Holocene CO_2 variations. Nevertheless, the observed Holocene CO_2 variations, combined with marine benthic $\delta^{13}C$ data (Shackleton 1977; Curry et al. 1988; Duplessy et al. 1988) are consistent with the timescale inferred for ocean-sediment interactions (Archer et al. 1999), and with a role for CO_2 fertilization in determining terrestrial carbon storage at concentrations of CO_2 within the natural range (180–280 ppm) (Esser and Lautenschlager 1994; Peng et al. 1998; Bennett and Willis 2000). One implication for the future evolution of atmospheric CO_2 is that the $CaCO_3$ compensation mechanism, which added CO_2 to the atmosphere during the preindustrial Holocene, will similarly remove atmospheric CO_2 emitted by human activities, but the timescale for this removal will be very long (multimillennia). The carbon sink mechanism due to CO_2 fertilization has additional experimental support for higher than present CO_2 concentrations (DeLucia et al. 1999; Luo et al. 1999), and is already accounted for in CO_2 stabilization scenario calculations (Joos et al. 2001; Prentice et al. 2001).

The Last Millennium

Knowledge of natural climate variability and of the role of solar and volcanic forcing is important for the attribution of climate change during recent decades and centuries. Different reconstructions of forcing by past solar irradiance variations, however, based on cosmogenic isotopes (^{14}C, ^{10}Be, ^{36}Cl; Beer et al. 1994) and sunspot numbers (Hoyt and Schatten 1994) vary by up to a factor of five (Lean et al. 1995; Reid 1997; Bard et al. 2000; Crowley 2000). In addition, there is a growing literature arguing for an amplification of the climate impact of solar irradiance changes through a solar modulation of the cosmic ray flux, affecting, for example, low-level cloud cover on Earth, based on empirical correlations between climate change and solar activity over specific periods (e.g., Tinsley and Deen 1991; Shaviv and Veizer 2003). The proposed correlation, however, breaks down over the last years of the instrumental record, and cosmic ray flux and climate change proxies are unrelated for the period 20 to 60 ka BP, when very large changes in the cosmic ray flux (Laschamp event) occurred (Wagner et al. 2001).

The magnitude of Northern Hemisphere surface temperature variations is also debated (Jones et al. 1998; Mann et al. 1999; Briffa 2000; Crowley 2000; Huang et al. 2000; Beltrami 2002; Esper et al. 2002). For example, Esper et al. (2002) report that the decadal-average Northern Hemisphere surface temperature was about 1 K lower in the first half of the 17th century than at around year AD 1000 and the present climatological mean, whereas Mann et al. (1999) report that decadal-average temperatures varied within 0.4 K only.

Highly resolved ice-core CO_2 records of the past millennium provide a joint constraint on Northern Hemisphere temperature variability and the climate–carbon cycle feedback. Figure 7.4 illustrates the relationship between the variations of decadal to multidecadal Northern Hemisphere surface temperature, atmospheric CO_2 concentration, and the strength of the climate–carbon cycle feedback (as ppm CO_2 released per K temperature increase in the Northern Hemisphere). The strength of the feedback depends on several factors, including the change in solubility of CO_2 in seawater and the responses of productivity and heterotrophic respiration to temperature. The current best estimate of the actual CO_2 range during the past millennium (before the Industrial Revolution) is 6 ppm, based on emerging high-quality measurements (Siegenthaler and Monnin, pers. comm.). If we assume a climate–carbon cycle feedback of 12 ppm K^{-1}, as obtained for the coupled Bern CC-LPJ model (Gerber et al. 2003), the CO_2 range of 6 ppm then constrains the decadal-average Northern Hemisphere surface temperature variation to a range of 0.5 K during the past millennium (excluding the rise that has occurred during recent decades; Figure 7.4). Alternatively, if we accept the ranges of temperature variations reconstructed by Mann et al. (1999) and by Esper et al. (2002) as equally possible, then the CO_2 concentration range of 6 ppm constrains the climate–carbon cycle feedback to between 6 and 16 ppm K^{-1} (for global mean surface temperature changes of less than ~1°C). This implies that the 0.6 ± 0.2°C increase in

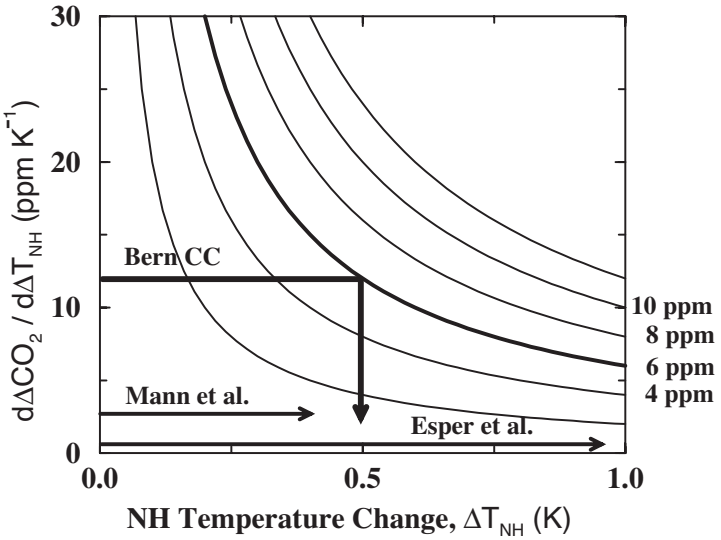

Figure 7.4. Relationship between Northern Hemisphere (NH) surface temperature change, climate–carbon cycle feedbacks, and variations in atmospheric CO_2. Isolines depict different ranges for CO_2 variation during the last millennium and are plotted against changes in decadal-average NH-temperature (horizontal axis) and the climate–carbon cycle feedback expressed as change in atmospheric CO_2 concentration per degree change in decadal-average NH surface temperature (vertical axis). The range of NH-temperature variations reconstructed by Mann et al. (1999) and Esper et al. (2002) are shown. The arrow labeled Bern CC depicts the NH temperature change obtained when combining the climate–carbon cycle feedback of the Bern CC-LPJ model (Gerber et al. 2003) with a CO_2 variability range of 6 ppm.

global mean surface temperature over the 20th century contributed little to the observed CO_2 increase over the same period. This data-based finding is somewhat in contrast with the model results of Dai and Fung (1993), which suggest climate change as a major driver of the terrestrial sink during the past century, but it is consistent with more recent model estimates that suggest a relatively small role of climate change in modulating the terrestrial sink over the past century (McGuire et al. 2001).

The temporal evolution of atmospheric CO_2 and Northern Hemisphere temperature in response to the reconstructed variability in solar irradiance and radiative forcing by volcanoes has been examined (Gerber et al. 2003). Modeled variations in atmospheric CO_2 and Northern Hemisphere mean surface temperature are compatible with reconstructions from different Antarctic ice cores and temperature proxy data for low solar forcing. Simulations where the magnitude of solar irradiance changes is increased toward the high estimates yield a mismatch between model results and the CO_2 meas-

urements. This finding provides further evidence for modest changes in solar irradiance and global mean temperature over the past millennium and argues against a substantial amplification of the response of global or hemispheric annual mean temperature to solar changes by the suggested cosmic ray flux–cloud cover link or other solar-related mechanisms. The results imply that solar changes are not the dominant factor in the 20th-century warming.

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