

Target Atmospheric CO₂: Where Should Humanity Aim?

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Abstract: Paleoclimate data show that climate sensitivity is ~3°C for doubled CO₂, including only fast feedback processes. Equilibrium sensitivity, including slower surface albedo feedbacks, is ~6°C for doubled CO₂ for the range of climate states between glacial conditions and ice-free Antarctica. Decreasing CO₂ was the main cause of a cooling trend that began 50 million years ago, large scale glaciation occurring when CO₂ fell to 450 ± 100 ppm, a level that will be exceeded within decades, barring prompt policy changes. If humanity wishes to preserve a planet similar to that on which civilization developed and to which life on Earth is adapted, paleoclimate evidence and ongoing climate change suggest that CO₂ will need to be reduced from its current 385 ppm to at most 350 ppm. The largest uncertainty in the target arises from possible changes of non-CO₂ forcings. An initial 350 ppm CO₂ target may be achievable by phasing out coal use except where CO₂ is captured and adopting agricultural and forestry practices that sequester carbon. If the present overshoot of this target CO₂ is not brief, there is a possibility of seeding irreversible catastrophic effects.

Keywords: climate change, climate sensitivity, global warming

1. INTRODUCTION

Human activities are altering Earth's atmospheric composition. Concern about global warming due to long-lived human-made greenhouse gases (GHGs) led to the United Nations Framework Convention on Climate Change [1] with the objective of stabilizing GHGs in the atmosphere at a level preventing "dangerous anthropogenic interference with the climate system."

The Intergovernmental Panel on Climate Change [IPCC, 2] and others [3] used several "reasons for concern" to estimate that global warming of more than 2-3°C may be dangerous. The European Union adopted 2°C above pre-industrial global temperature as a goal to limit human-made warming [4]. Hansen et al. [5] argued for a limit of 1°C global warming (relative to 2000, 1.7°C relative to pre-industrial time), aiming to avoid practically irreversible ice sheet and species loss. This 1°C limit, with nominal climate sensitivity of 3/4°C per W/m² and plausible control of other GHGs [6], implies maximum CO₂ ~ 450 ppm [5].

Our current analysis suggests that humanity must aim for an even lower level of GHGs. Paleoclimate data and ongoing global changes indicate that 'slow' climate feedback processes not included in most climate models, such as ice sheet disintegration, vegetation migration, and

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GHG release from soils, tundra or ocean sediments, may begin to come into play on time scales as short as centuries or less [7]. Rapid on-going climate changes and realization that Earth is out of energy balance, implying that more warming is ‘in the pipeline’ [8], add urgency to investigation of the dangerous level of GHGs.

A probabilistic analysis [9] concluded that the long-term CO₂ limit is in the range 300-500 ppm for 25 percent risk tolerance, depending on climate sensitivity and non-CO₂ forcings. Stabilizing atmospheric CO₂ and climate requires that net CO₂ emissions approach zero, because of the long lifetime of CO₂ [10, 11].

We use paleoclimate data to show that long-term climate has high sensitivity to climate forcings and that the present global mean CO₂, 385 ppm, is already in the dangerous zone. Despite rapid current CO₂ growth, ~2 ppm/year, we show that it is conceivable to reduce CO₂ this century to less than the current amount, but only via prompt policy changes.

1.1. Climate sensitivity

A global climate forcing, measured in W/m² averaged over the planet, is an imposed perturbation of the planet’s energy balance. Increase of solar irradiance (S₀) by 2% and doubling of atmospheric CO₂ are each forcings of about 4 W/m² [12].

Charney [13] defined an idealized climate sensitivity problem, asking how much global surface temperature would increase if atmospheric CO₂ were instantly doubled, assuming that slowly-changing planetary surface conditions, such as ice sheets and forest cover, were fixed. Long-lived GHGs, except for the specified CO₂ change, were also fixed, not responding to climate change. The Charney problem thus provides a measure of climate sensitivity including only the effect of ‘fast’ feedback processes, such as changes of water vapor, clouds and sea ice.

Classification of climate change mechanisms into fast and slow feedbacks is useful, even though time scales of these changes may overlap. We include as fast feedbacks aerosol changes, e.g., of desert dust and marine dimethylsulfide, that occur in response to climate change [7].

Charney [13] used climate models to estimate fast-feedback doubled CO₂ sensitivity of 3 ± 1.5°C. Water vapor increase and sea ice decrease in response to global warming were both found to be strong positive feedbacks, amplifying the surface temperature response. Climate models in the current IPCC [2] assessment still agree with Charney’s estimate.

Climate models alone are unable to define climate sensitivity more precisely, because it is difficult to prove that models realistically incorporate all feedback processes. The Earth’s history, however, allows empirical inference of both fast feedback climate sensitivity and long-term sensitivity to specified GHG change including the slow ice sheet feedback.

2. PLEISTOCENE EPOCH

Atmospheric composition and surface properties in the late Pleistocene are known well enough for accurate assessment of the fast-feedback (Charney) climate sensitivity. We first compare the pre-industrial Holocene with the last glacial maximum [LGM, 20 ky BP (before present)]. The planet was in energy balance in both periods within a small fraction of 1 W/m², as shown by considering the contrary: an imbalance of 1 W/m² maintained a few millennia would melt all ice on the planet or change ocean temperature an amount far outside measured variations [Table S1 of 8]. The approximate equilibrium characterizing most of Earth’s history is unlike the current situation, in which GHGs are rising at a rate much faster than the coupled climate system can respond.

Climate forcing in the LGM equilibrium state due to the ice age surface properties, i.e., increased ice area, different vegetation distribution, and continental shelf exposure, was $-3.5 \pm 1 \text{ W/m}^2$ [14] relative to the Holocene. Additional forcing due to reduced amounts of long-lived GHGs (CO_2 , CH_4 , N_2O), including the indirect effects of CH_4 on tropospheric ozone and stratospheric water vapor (Fig. S1) was $-3 \pm 0.5 \text{ W/m}^2$. Global forcing due to slight changes in the Earth's orbit is a negligible fraction of 1 W/m^2 (Fig. S2). The total 6.5 W/m^2 forcing and global surface temperature change of $5 \pm 1^\circ\text{C}$ relative to the Holocene [15, 16] yield an empirical sensitivity $\sim 3/4 \pm 1/4^\circ\text{C}$ per W/m^2 forcing, i.e., a Charney sensitivity of $3 \pm 1^\circ\text{C}$ for the 4 W/m^2 forcing of doubled CO_2 . This empirical fast-feedback climate sensitivity allows water vapor, clouds, aerosols, sea ice, and all other fast feedbacks that exist in the real world to respond naturally to global climate change.

Climate sensitivity varies as Earth becomes warmer or cooler. Toward colder extremes, as the area of sea ice grows, the planet approaches runaway snowball-Earth conditions, and at high temperatures it can approach a runaway greenhouse effect [12]. At its present temperature Earth is on a flat portion of its fast-feedback climate sensitivity curve (Fig. S3). Thus our empirical sensitivity, although strictly the mean fast-feedback sensitivity for climate states ranging from the ice age to the current interglacial period, is also today's fast-feedback climate sensitivity.

2.1. Verification

Our empirical fast-feedback climate sensitivity, derived by comparing conditions at two points in time, can be checked over the longer period of ice core data. Fig. (1a) shows CO_2 and CH_4 data from the Antarctic Vostok ice core [17, 18] and sea level based on Red Sea sediment cores [18]. Gases are from the same ice core and have a consistent time scale, but dating with respect to sea level may have errors up to several thousand years.

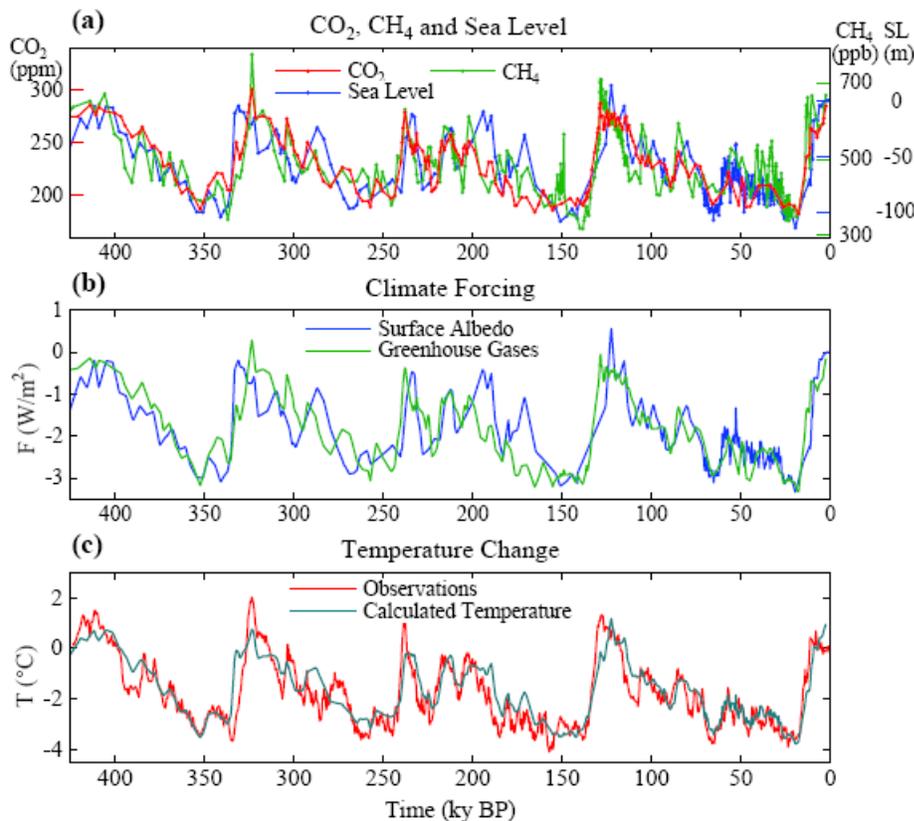


Fig. (1). (a) CO_2 , CH_4 [17] and sea level [19] for past 425 ky.

(b) Climate forcings due to changes of GHGs and ice sheet area, the latter inferred from sea level change.

(c) Calculated global temperature change based on climate sensitivity of $3/4^\circ\text{C}$ per W/m^2 . Observations are Antarctic temperature change [18] divided by two.

We use the GHG and sea level data to calculate climate forcing by GHGs and surface albedo change as in prior calculations [7], but with two refinements. First, we specify the N₂O climate forcing as 12 percent of the sum of the CO₂ and CH₄ forcings, rather than the 15 percent estimated earlier [7]. Because N₂O data are not available for the entire record, and its forcing is small and highly correlated with CO₂ and CH₄, we take the GHG effective forcing as

$$F_e(\text{GHGs}) = 1.12 [F_a(\text{CO}_2) + 1.4 F_a(\text{CH}_4)], \quad (1)$$

using published formulae for F_a of each gas [20]. The factor 1.4 accounts for the higher efficacy of CH₄ relative to CO₂, which is due mainly to the indirect effect of CH₄ on tropospheric ozone and stratospheric water vapor [12]. The resulting GHG forcing between the LGM and late Holocene is 3 W/m², apportioned as 75% CO₂, 14% CH₄ and 11% N₂O.

The second refinement in our calculations is to surface albedo. Based on models of ice sheet shape, we take the horizontal area of the ice sheet as proportional to the 4/5 power of volume. Fig. (S4) compares our present albedo forcing with prior use [7] of exponent 2/3, showing that this choice and division of the ice into multiple ice sheets has only a minor effect.

Multiplying the sum of GHG and surface albedo forcings by climate sensitivity $\frac{3}{4}^\circ\text{C}$ per W/m² yields the blue curve in Fig. (1c). Vostok temperature change [17] divided by two (red curve) is used to crudely estimate global temperature change, as typical glacial-interglacial global annual-mean temperature change is $\sim 5^\circ\text{C}$ and is associated with $\sim 10^\circ\text{C}$ change on Antarctica [21]. Fig. (1c) shows that fast-feedback climate sensitivity $\frac{3}{4}^\circ\text{C}$ per W/m² (3°C for doubled CO₂) is a good approximation for the entire period.

2.2. Slow feedbacks

Let us consider climate change averaged over a few thousand years – long enough to assure energy balance and minimize effects of ocean thermal response time and climate change leads/lags between hemispheres [22]. At such temporal resolution the temperature variations in Fig. (1) are global, with high latitude amplification, being present in sea surface temperature derived from ocean sediment cores and polar ice cores (Fig. S5).

GHG and surface albedo changes are mechanisms causing the large global climate changes in Fig. (1), but they do not initiate these climate swings. Instead changes of GHGs and sea level (a measure of ice sheet size) lag temperature change by several hundred years [6, 7, 23, 24].

GHG and surface albedo changes are positive climate feedbacks. Major glacial-interglacial climate swings are instigated by slow changes of Earth's orbit, especially the tilt of Earth's spin-axis relative to the orbital plane and the precession of the equinoxes that influences the intensity of summer insolation [25, 26]. Global radiative forcing due to orbital changes is small, but ice sheet size is affected by changes of geographical and seasonal insolation (e.g., ice melts at both poles when the spin-axis tilt increases, and ice melts at one pole when perihelion, the closest approach to the sun, occurs in late spring [7]). Also a warming climate causes net release of GHGs. The most effective GHG feedback is release of CO₂ by the ocean, due partly to temperature dependence of CO₂ solubility but mostly to increased ocean mixing in a warmer climate, which acts to flush out deep ocean CO₂ and alters ocean biological productivity [27].

GHG and surface albedo feedbacks respond and contribute to temperature change caused by any climate forcing, natural or human-made, given sufficient time. The GHG feedback is nearly linear in global temperature during the late Pleistocene [Fig. (7) of 6, 28]. Surface albedo feedback increases as Earth becomes colder and the area of ice increases. Climate sensitivity on

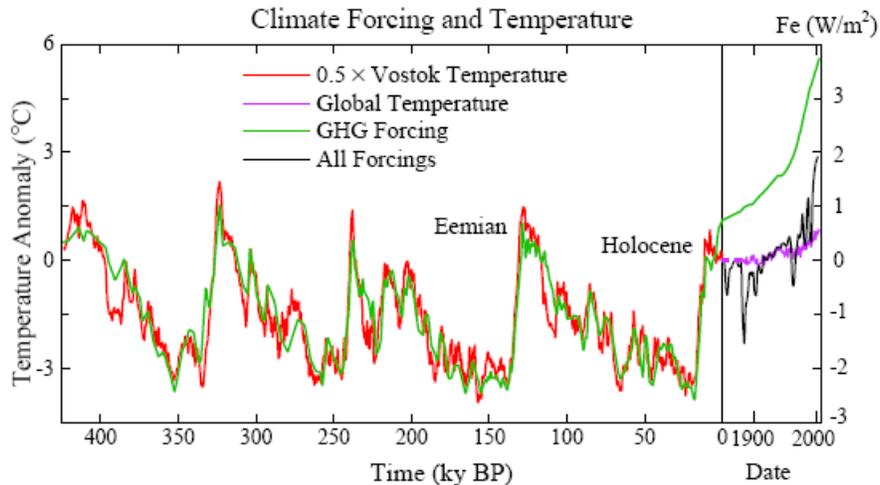


Fig. (2). Global temperature (left scale) and GHG forcing (right scale) due to CO₂, CH₄ and N₂O from the Vostok ice core [17, 18]. Ttime scale is expanded for the industrial era. Ratio of temperature and forcing scales is 1.5°C per W/m², i.e., the temperature scale gives the expected equilibrium response to GHG change including (slow feedback) surface albedo change. Modern forcings include human-made aerosols, volcanic aerosols and solar irradiance [5]. GHG forcing zero point is the mean for 10-8 ky BP (Fig. S6). Zero point of modern temperature and net climate forcing was set at 1850 [5], but this is also the zero point for 10-8 ky BP, as shown by the absence of a trend in Fig. (S6) and by the discussion of that figure.

Pleistocene time scales includes slow feedbacks, and is larger than the Charney sensitivity estimate, because the dominant slow feedbacks are positive. Other feedbacks, e.g., increased weathering as CO₂ increases, become important on longer geologic time scales.

Paleoclimate data permit evaluation of long-term sensitivity to specified GHG change. We assume only that the area of ice is a function of global temperature. Plotting GHG forcing [7] from ice core data [18] against temperature shows that global climate sensitivity including the slow surface albedo feedback is 1.5°C per W/m² or 6°C for doubled CO₂ (Fig. 2), twice as large as the Charney fast-feedback sensitivity.

This long-term climate sensitivity is relevant to GHGs that remain airborne for centuries-to-millennia. The human-caused atmospheric GHG increase will decline slowly if anthropogenic emissions from fossil fuel burning decrease enough, as we illustrate below using a simplified carbon cycle model. On the other hand, if the globe warms much further, carbon cycle models [2] and empirical data [6, 28] reveal a positive GHG feedback on century-millennia time scales. This amplification of GHG amount is moderate if warming is kept within the range of recent interglacial periods [6], but larger warming would risk greater release of CH₄ and CO₂ from methane hydrates in tundra and ocean sediments [29]. On still longer, geological, time scales weathering of rocks causes a negative feedback on atmospheric CO₂ amount [30], as discussed in section 3, but this feedback is too slow to alleviate climate change of concern to humanity.

2.3 Time scales

How long does it take to reach equilibrium temperature? Response is slowed by ocean thermal inertia and the time needed for ice sheets to disintegrate.

Ocean-caused delay is estimated in Fig. (S7) using a coupled atmosphere-ocean model. One-third of the response occurs in the first few years, in part because of rapid response over land, one-half in ~25 years, three-quarters in 250 years, and nearly full response in a millennium. The ocean-caused delay is a strong (quadratic) function of climate sensitivity and it depends on the rate of mixing of surface water and deep water [31], as discussed in the Appendix.

Ice sheet response time is often assumed to be several millennia, based on the broad sweep of paleo sea level change (Fig. 1a) and primitive ice sheet models designed to capture that change. However, this long time scale may reflect the slowly changing orbital forcing, rather than inherent inertia, as there is no discernable lag between maximum ice sheet melt rate and local insolation that favors melt [7]. Paleo sea level data with high time resolution reveal frequent ‘suborbital’ sea level changes at rates of 1 m/century or more [32-34].

Present-day observations of Greenland and Antarctica show increasing surface melt [35], loss of buttressing ice shelves [36], accelerating ice streams [37], and increasing overall mass loss [38]. These rapid changes do not occur in existing ice sheet models, which are missing critical physics of ice sheet disintegration [39]. Sea level changes of several meters per century occur in the paleoclimate record [32, 33], in response to forcings slower and weaker than the present human-made forcing. It seems likely that large ice sheet response will occur within centuries, if human-made forcings continue to increase. Once ice sheet disintegration is underway, decadal changes of sea level may be substantial.

2.4. Warming “in the pipeline”

The expanded time scale for the industrial era (Fig. 2) reveals a growing gap between actual global temperature (purple curve) and equilibrium (long-term) temperature response based on the net estimated climate forcing (black curve). Ocean and ice sheet response times together account for this gap, which is now 2.0°C.

The forcing in Fig. (2) (black curve, Fe scale), when used to drive a global climate model [5], yields global temperature change that agrees closely [Fig. (3) in 5] with observations (purple curve, Fig. 2). That climate model, which includes only fast feedbacks, has additional warming of ~0.6°C in the pipeline today because of ocean thermal inertia [5, 8].

The remaining gap between equilibrium temperature for current atmospheric composition and actual global temperature is ~1.4°C. This further 1.4°C warming still to come is due to the slow surface albedo feedback, specifically ice sheet disintegration and vegetation change.

One may ask whether the climate system, as the Earth warms from its present ‘interglacial’ state, still has the capacity to supply slow feedbacks that double the fast-feedback sensitivity. This issue can be addressed by considering longer time scales including periods with no ice.

3. CENOZOIC ERA

Pleistocene atmospheric CO₂ variations occur as a climate feedback, as carbon is exchanged among its surface reservoirs: the ocean, atmosphere, soils and biosphere. The most effective feedback is the increase of atmospheric CO₂ as climate warms, the CO₂ transfer being primarily from the ocean to the atmosphere [27, 28]. On longer time scales the total amount of CO₂ in the surface reservoirs can change due to exchange of carbon with the solid earth. CO₂ thus becomes a primary agent of long-term climate change, leaving orbital effects as ‘noise’ on larger climate swings.

The Cenozoic era, the past 65.5 My, provides a valuable complement to the Pleistocene for exploring climate sensitivity. Cenozoic data on climate and atmospheric composition are not as precise, but larger climate variations occur, including an ice-free planet, thus putting glacial-interglacial changes in a wider perspective.

Oxygen isotopic composition of benthic (deep ocean dwelling) foraminifera shells in a global compilation of ocean sediment cores [26] provides a starting point for analyzing Cenozoic climate change (Fig. 3a). At times with negligible ice sheets, oxygen isotope change, $\delta^{18}\text{O}$,

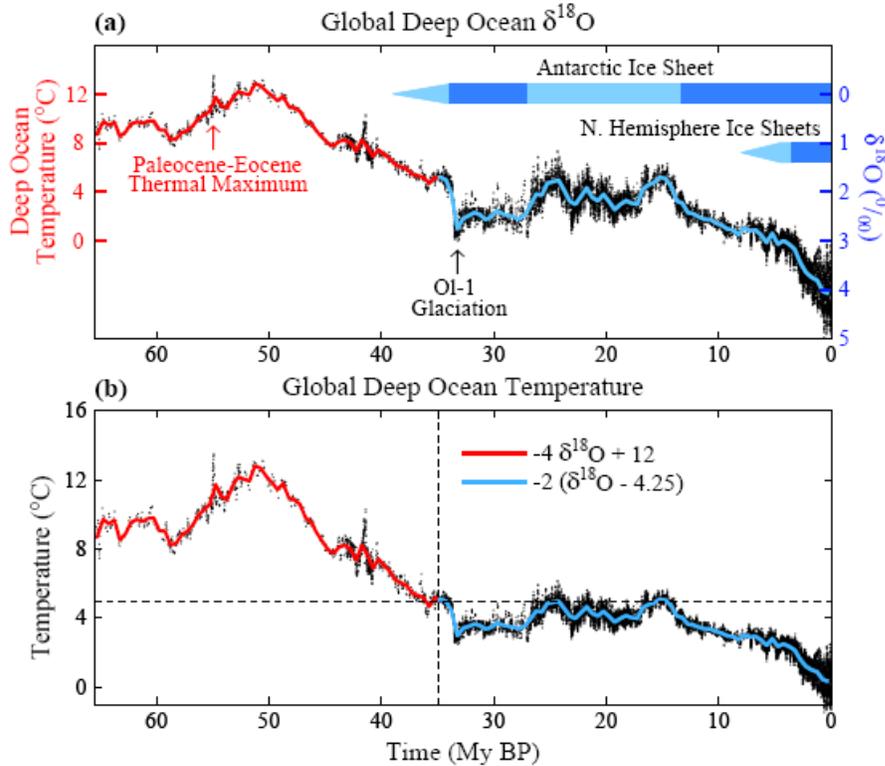


Fig. (3). Global deep ocean (a) $\delta^{18}\text{O}$ [26] and (b) temperature. Black curve is 5-point running mean of $\delta^{18}\text{O}$ original temporal resolution, while red and blue curves have 500 ky resolution.

provides a direct measure of deep ocean temperature (T_{do}). Thus T_{do} ($^{\circ}\text{C}$) $\sim -4 \delta^{18}\text{O} + 12$ between 65.5 and 34 My BP.

Rapid increase of $\delta^{18}\text{O}$ at about 34 My is associated with glaciation of Antarctica [26, 40] and global cooling, as evidenced by data from North America [41] and Asia [42]. From then until the present, ^{18}O in deep ocean foraminifera is affected by both ice volume and T_{do} , lighter ^{16}O evaporating preferentially from the ocean and accumulating in ice sheets. Between 34 My and the last ice age (20 ky) the change of $\delta^{18}\text{O}$ was $\sim 3\text{‰}$, change of T_{do} was $\sim 6^{\circ}\text{C}$ (from $+5$ to -1°C) and ice volume change ~ 180 msl (meters of sea level). Given that a 1.5‰ change of $\delta^{18}\text{O}$ is associated with a 6°C T_{do} change, we assign the remaining $\delta^{18}\text{O}$ change to ice volume linearly at the rate 60 msl per mil $\delta^{18}\text{O}$ change (thus 180 msl for $\delta^{18}\text{O}$ between 1.75 and 4.75). Equal division of $\delta^{18}\text{O}$ between temperature and sea level yields sea level change in the late Pleistocene in reasonable accord with available sea level data (Fig. S8). Subtracting the ice volume portion of $\delta^{18}\text{O}$ yields deep ocean temperature T_{do} ($^{\circ}\text{C}$) $= -2(\delta^{18}\text{O} - 4.25\text{‰})$ after 34 My, as in Fig. (3b).

The large ($\sim 14^{\circ}\text{C}$) Cenozoic temperature change between 50 My and the ice age at 20 ky must have been forced by changes of atmospheric composition. Alternative drives could come from outside (solar irradiance) or the Earth's surface (continental locations). But solar brightness increased $\sim 0.4\%$ in the Cenozoic [43], a linear forcing change of only $+1 \text{ W/m}^2$ and of the wrong sign to contribute to the cooling trend. Climate forcing due to continental locations was $< 1 \text{ W/m}^2$, because continents 65 My ago were already close to present latitudes (Fig. S9). Opening or closing of oceanic gateways might affect the timing of glaciation, but it would not provide the climate forcing needed for global cooling.

CO_2 concentration, in contrast, varied from ~ 180 ppm in glacial times to 1500 ± 500 ppm in the early Cenozoic [44]. This change is a forcing of more than 10 W/m^2 (Table 1 in [16]), an order of magnitude larger than other known forcings. CH_4 and N_2O , positively correlated with

CO₂ and global temperature in the period with accurate data (ice cores), likely increase the total GHG forcing, but their forcings are much smaller than that of CO₂ [45, 46].

3.1. Cenozoic carbon cycle

Solid Earth sources and sinks of CO₂ are not, in general, balanced at any given time [30, 47]. CO₂ is removed from surface reservoirs by: (1) chemical weathering of rocks with deposition of carbonates on the ocean floor, and (2) burial of organic matter; weathering is the dominant process [30]. CO₂ returns primarily via metamorphism and volcanic outgassing where carbonate-rich oceanic crust is subducted beneath moving continental plates.

Outgassing and burial of CO₂ are each typically 10¹²-10¹³ mol C/year [30, 47-48]. At times of unusual plate tectonic activity, such as rapid subduction of carbon-rich ocean crust or strong orogeny, the imbalance between outgassing and burial can be a significant fraction of the one-way carbon flux. Although negative feedbacks in the geochemical carbon cycle reduce the rate of surface reservoir perturbation [49], a net imbalance ~10¹² mol C/year can be maintained over thousands of years. Such an imbalance, if confined to the atmosphere, would be ~0.005 ppm/year, but as CO₂ is distributed among surface reservoirs, this is only ~0.0001 ppm/year. This rate is negligible compared to the present human-made atmospheric CO₂ increase of ~2 ppm/year, yet over a million years such a crustal imbalance alters atmospheric CO₂ by 100 ppm.

Between 60 and 50 My ago India moved north rapidly, 18-20 cm/year [50], through a region that long had been a depocenter for carbonate and organic sediments. Subduction of carbon-rich crust was surely a large source of CO₂ outgassing and a prime cause of global warming, which peaked 50 My ago (Fig. 3b) with the Indo-Asian collision. CO₂ must have then decreased due to a reduced subduction source and enhanced weathering with uplift of the Himalayas/Tibetan Plateau [51]. Since that time the Indian and Atlantic Oceans have been the major depocenters for carbon, with subduction of carbon-rich crust limited mainly to small regions near Indonesia and Central America [47].

Thus atmospheric CO₂ declined following the Indo-Asian collision [44] and climate cooled (Fig. 3b) leading to Antarctic glaciation by ~34 My. Antarctica has been more or less glaciated ever since. The rate of CO₂ drawdown declines as atmospheric CO₂ decreases due to negative feedbacks, including the effect of declining atmospheric temperature and plant growth rates on weathering [30]. These negative feedbacks tend to create a balance between crustal outgassing and drawdown of CO₂, which have been equal within 1-2 percent over the past 700 ky [52]. Large fluctuations in the size of the Antarctic ice sheet have occurred, possibly related to temporal variations of plate tectonics [53] and outgassing rates. The relatively constant atmospheric CO₂ amount of the past 20 My (Fig. S10) implies a near balance of global outgassing and weather rates over that period.

Knowledge of Cenozoic CO₂ is limited to imprecise proxy measures except for recent ice core data. There are discrepancies among different proxy measures, and even between different investigators using the same proxy method, as discussed in conjunction with Fig. (S10). Nevertheless, the proxy data indicate that CO₂ was of the order of 1000 ppm in the early Cenozoic but <500 ppm in the last 20 My [2, 44].

3.2. Cenozoic forcing and CO₂

The entire Cenozoic climate forcing history (Fig. 4a) is implied by the temperature reconstruction (Fig. 3b), assuming a fast-feedback sensitivity of ¾°C per W/m². Subtracting the solar and surface albedo forcings (Fig. 4b), the latter from Eq. S2 with ice sheet area vs. time from δ¹⁸O, we obtain the GHG forcing history (Fig. 4c).

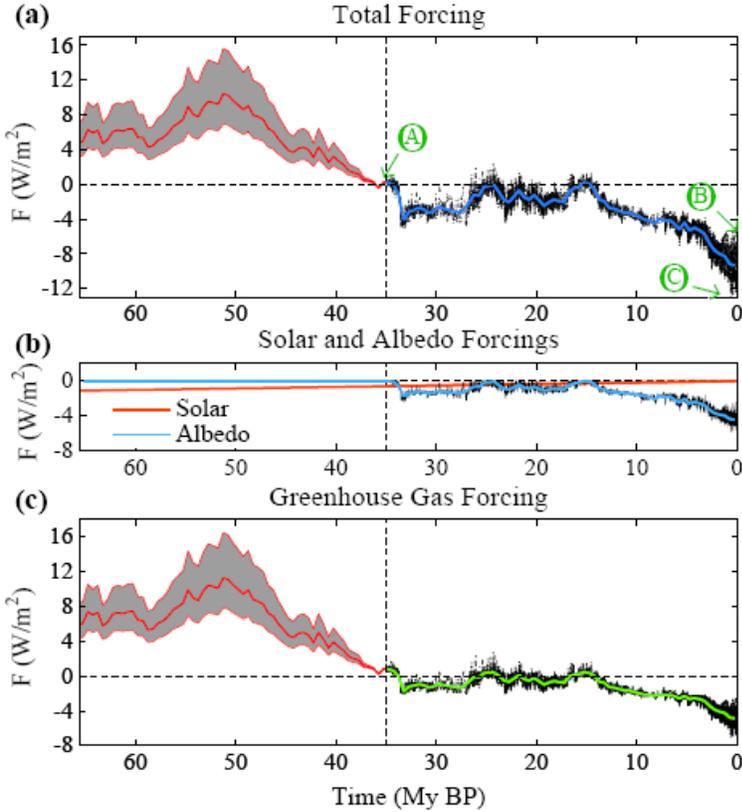


Fig. (4). (a) Total climate forcing, (b) solar and surface albedo forcings, and (c) GHG forcing in the Cenozoic, based on T_{do} history of Fig. (3b) and assumed fast-feedback climate sensitivity $\frac{3}{4}^{\circ}\text{C}$ per W/m^2 . Ratio of T_s change and T_{do} change is assumed to be near unity in the minimal ice world between 65 and 35 My, but the gray area allows for 50% uncertainty in the ratio. In the later era with large ice sheets we take $\Delta T_s/\Delta T_{do} = 1.5$, in accord with Pleistocene data.

We hinge our calculations at 35 My for several reasons. Between 65 and 35 My ago there was little ice on the planet, so climate sensitivity is defined mainly by fast feedbacks. Second, we want to estimate the CO_2 amount that precipitated Antarctic glaciation. Finally, the relation between global surface air temperature change (ΔT_s) and deep ocean temperature change (ΔT_{do}) differs for ice-free and glaciated worlds.

Climate models show that global temperature change is tied closely to ocean temperature change [54]. Deep ocean temperature is a function of high latitude ocean surface temperature, which tends to be amplified relative to global mean ocean surface temperature. However, land temperature change exceeds that of the ocean, with an effect on global temperature that tends to offset the latitudinal variation of ocean temperature. Thus in the ice-free world (65-35 My) we take $\Delta T_s \sim \Delta T_{do}$ with generous (50%) uncertainty. In the glaciated world ΔT_{do} is limited by the freezing point in the deep ocean. ΔT_s between the last ice age (20 ky) and the present interglacial period ($\sim 5^{\circ}\text{C}$) was ~ 1.5 times larger than ΔT_{do} . In Fig. (S5) we show that this relationship fits well throughout the period of ice core data.

If we specify CO_2 at 35 My, the GHG forcing defines CO_2 at other times, assuming CO_2 provides 75% of the GHG forcing, as in the late Pleistocene. $\text{CO}_2 \sim 450$ ppm at 35 My keeps CO_2 in the range of early Cenozoic proxies (Fig. 5a) and yields a good fit to the amplitude and mean CO_2 amount in the late Pleistocene (Fig. 5b). A ~ 500 ppm CO_2 threshold for Antarctic glaciation was previously inferred from proxy CO_2 data and a carbon cycle model [55].

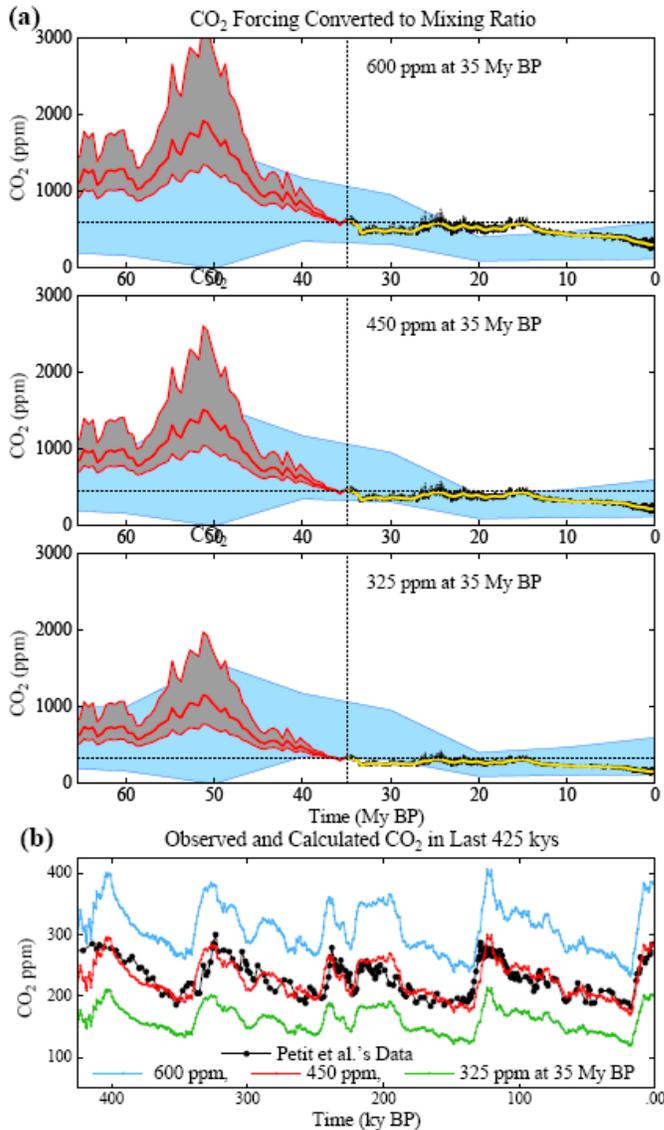


Fig. (5). (a) Simulated CO₂ amounts in the Cenozoic for three choices of CO₂ amount at 35 My (temporal resolution of black and colored curves as in Fig. (3); blue region: multiple CO₂ proxy data, discussed with Fig. S10; gray region allows 50 percent uncertainty in ratio of global surface and deep ocean temperatures). (b) Expanded view of late Pleistocene, including precise ice core CO₂ measurements (black curve).

Individual CO₂ proxies (Fig. S10) clarify limitations due to scatter among the measurements. Low CO₂ of some early Cenozoic proxies, if valid, would suggest higher climate sensitivity. However, in general the sensitivities inferred from the Cenozoic and Phanerozoic [56, 57, 58] agree well with our analysis, if we account for the ways in which sensitivity is defined and the periods emphasized in each empirical derivation (Table S1).

Our CO₂ estimate of ~450 ppm at 35 My (Fig. 5) serves as a prediction to compare with new data on CO₂ amount. Model uncertainties (Fig. S10) include possible changes of non-CO₂ GHGs and the relation of ΔT_s to ΔT_{do} . The model fails to account for cooling in the past 15 My if CO₂ increased, as several proxies suggest (Fig. S10). Changing ocean currents, such as the closing of the Isthmus of Panama, may have contributed to climate evolution, but models find little effect on temperature [59]. Non-CO₂ GHGs also could have played a role, because little forcing would have been needed to cause cooling due to the magnitude of late Cenozoic albedo feedback.

3.3. Implication

We infer from Cenozoic data that CO₂ was the dominant Cenozoic forcing, that CO₂ was ~450 ± 100 ppm when Antarctica glaciated, and that glaciation is reversible. Together these inferences have profound implications.

Consider three points marked in Fig. (4): point A at 35 My, just before Antarctica glaciated; point B at recent interglacial periods; point C at the depth of recent ice ages. Point B is about half way between A and C in global temperature (Fig. 3b) and climate forcings (Fig. 4). The GHG forcing from the deepest recent ice age to current interglacial warmth is ~3.5 W/m². Additional 4 W/m² forcing carries the planet, at equilibrium, to the ice-free state. Thus equilibrium climate sensitivity to GHG change, including the surface albedo change as a slow feedback, is almost as large between today and an ice-free world as between today and the ice ages.

The implication is that global climate sensitivity of 3°C for doubled CO₂, although valid for the idealized Charney definition of climate sensitivity, is a considerable understatement of expected equilibrium global warming in response to imposed doubled CO₂. Additional warming, due to slow climate feedbacks including loss of ice and spread of flora over the vast high-latitude land area in the Northern Hemisphere, approximately doubles equilibrium climate sensitivity.

The inferred high climate sensitivity between today's climate and the ice-free world is partly dependent on our estimate that global surface air temperature exceeded deep ocean temperature change by a factor 1.5 in the glaciated world. On the other hand, the GHGs tend to increase in response to global warming, and even if we include the positive CO₂ feedback in specifying an imposed CO₂ forcing, non-CO₂ GHG positive feedback adds to the climate response. Thus we conclude that our estimate that equilibrium climate sensitivity to doubled CO₂ including slow feedbacks is about twice the idealized Charney sensitivity is unlikely to be an exaggeration.

Equilibrium climate response would not be reached in decades or even in a century, because surface warming is slowed by the inertia of the ocean (Fig. S7) and ice sheets. However, Earth's history shows that the positive feedbacks allow global warmings to be relatively rapid, including sea level rise as fast as several meters per century [7]. Thus if humans push the climate system sufficiently far into disequilibrium, positive climate feedbacks may set in motion dramatic climate change and climate impacts that cannot be controlled.

4. ANTHROPOCENE ERA

Human-made global climate forcings now prevail over natural forcings (Fig. 2). Earth may have entered the Anthropocene era [60, 61] 6-8 ky ago [62], but the net human-made forcing was small, perhaps slightly negative [7], prior to the industrial era. GHG forcing overwhelmed natural and negative human-made forcings only in the past quarter century (Fig. 2).

Human-made climate change is delayed by ocean (Fig. S7) and ice sheet response times. Warming 'in the pipeline', mostly attributable to slow feedbacks, is now about 2°C (Fig. 2). No additional forcing is required to raise global temperature to at least the level of the Pliocene, 2-3 million years ago, a degree of warming that would surely yield 'dangerous' climate impacts [5].

4.1. Tipping points

Realization that today's climate is far out of equilibrium with current climate forcings raises the specter of 'tipping points', the concept that climate can reach a point where, without additional forcing, rapid changes proceed practically out of our control [2, 7, 63]. Arctic sea ice and the West Antarctic Ice Sheet are examples of potential tipping points. Arctic sea ice loss is magnified by the positive feedback of increased absorption of sunlight as global warming

initiates sea ice retreat [64]. West Antarctic ice loss can be accelerated by several feedbacks, once ice loss is substantial [39].

We define: (1) the *tipping level*, the global climate forcing that, if long maintained, gives rise to a specific consequence, and (2) the *point of no return*, a climate state beyond which the consequence is inevitable, even if climate forcings are reduced. A point of no return can be avoided, even if the tipping level is temporarily exceeded. Ocean and ice sheet inertia permit overshoot, provided the climate forcing is returned below the tipping level before initiating irreversible dynamic change.

Points of no return are inherently difficult to define, because the dynamical problems are nonlinear. Existing models are more lethargic than the real world for phenomena now unfolding, including changes of sea ice [65], ice streams [66], ice shelves [36], and expansion of the subtropics [67, 68].

The tipping level is easier to assess, because the paleoclimate equilibrium response to known climate forcing is relevant. The tipping level is a measure of the long-term climate forcing that humanity must aim to stay beneath to avoid large climate impacts. The tipping level does not define the magnitude or period of tolerable overshoot. However, if overshoot is in place for centuries, the thermal perturbation will so penetrate the ocean [10] that recovery without dramatic effects, such as ice sheet disintegration, becomes unlikely.

4.2. Target CO₂

GHGs other than CO₂ cause climate forcing comparable to that of CO₂ [2, 6], but growth of non-CO₂ GHGs is falling below IPCC [2] scenarios and the total GHG climate forcing change is determined mainly by CO₂ [69]. Net human-made forcing is comparable to the CO₂ forcing, as non-CO₂ GHGs tend to offset negative aerosol forcing [2, 5].

Thus we take future CO₂ change as approximating the net human-made forcing change, with two caveats. First, special effort to reduce non-CO₂ GHGs could alleviate the CO₂ requirement, allowing up to about +25 ppm CO₂ for the same climate effect, while resurgent growth of non-CO₂ GHGs could reduce allowed CO₂ a similar amount [6]. Second, reduction of human-made aerosols, which have a net cooling effect, could force stricter GHG requirements. However, an emphasis on reducing black soot could largely off-set reductions of high albedo aerosols [20].

Our estimated history of CO₂ through the Cenozoic Era provides a sobering perspective for assessing an appropriate target for future CO₂ levels. A CO₂ amount of order 450 ppm or larger, if long maintained, would push Earth toward the ice-free state. Although ocean and ice sheet inertia limit the rate of climate change, such a CO₂ level likely would cause the passing of climate tipping points and initiate dynamic responses that could be out of humanity's control.

The climate system, because of its inertia, has not yet fully responded to the recent increase of human-made climate forcings [5]. Yet climate impacts are already occurring that allow us to make an initial estimate for a target atmospheric CO₂ level. No doubt the target will need to be adjusted as climate data and knowledge improve, but the urgency and difficulty of reducing the human-made forcing will be less, and more likely manageable, if excess forcing is limited soon.

Civilization is adapted to climate zones of the Holocene. Theory and models indicate that subtropical regions expand poleward with global warming [2, 67]. Data reveal a 4-degree latitudinal shift already [68], larger than model predictions, yielding increased aridity in southern United States [70, 71], the Mediterranean region, Australia and parts of Africa. Impacts of this climate shift [72] support the conclusion that 385 ppm CO₂ is already deleterious.

Alpine glaciers are in near-global retreat [72, 73]. After a flush of fresh water, glacier loss foretells long summers and autumns of frequently dry rivers, including rivers originating in the

Himalayas, Andes and Rocky Mountains that now supply water to hundreds of millions of people. Present glacier retreat, and warming in the pipeline, indicate that 385 ppm CO₂ is already a threat.

Equilibrium sea level rise for today's 385 ppm CO₂ is at least several meters, judging from paleoclimate history [19, 32-34]. Accelerating mass losses from Greenland [74] and West Antarctica [75] heighten concerns about ice sheet stability. An initial CO₂ target of 350 ppm, to be reassessed as effects on ice sheet mass balance are observed, is suggested.

Stabilization of Arctic sea ice cover requires, to first approximation, restoration of planetary energy balance. Climate models driven by known forcings yield a present planetary energy imbalance of +0.5-1 W/m² [5], a result supported by observed increasing ocean heat content [76]. CO₂ amount must be reduced to 325-355 ppm to increase outgoing flux 0.5-1 W/m², if other forcings are unchanged. A further reduced flux, by ~0.5 W/m², and thus CO₂ ~300-325 ppm, may be needed to restore sea ice to its area of 25 years ago.

Coral reefs are suffering from multiple stresses, with ocean acidification and ocean warming principal among them [77]. Given additional warming 'in-the-pipeline', 385 ppm CO₂ is already deleterious. A 300-350 ppm CO₂ target would significantly relieve both of these stresses.

4.3. CO₂ scenarios

A large fraction of fossil fuel CO₂ emissions stays in the air a long time, one-quarter remaining airborne for several centuries [11, 78, 79]. Thus moderate delay of fossil fuel use will not appreciably reduce long-term human-made climate change. Preservation of climate requires that most remaining fossil fuel carbon is never emitted to the atmosphere.

Coal is the largest reservoir of conventional fossil fuels (Fig. **S12**), exceeding combined reserves of oil and gas [2, 79]. The only realistic way to sharply curtail CO₂ emissions is to phase out coal use except where CO₂ is captured and sequestered.

Phase-out of coal emissions by 2030 (Fig. **6**) keeps maximum CO₂ close to 400 ppm, depending on oil and gas reserves and reserve growth. IPCC reserves assume that half of readily extractable oil has already been used (Figs. **6**, **S12**). EIA [80] estimates (Fig. **S12**) have larger reserves and reserve growth. Even if EIA estimates are accurate, the IPCC case remains valid if the most difficult to extract oil and gas is left in the ground, via a rising price on carbon emissions that discourages remote exploration and environmental regulations that place some areas off-limit. If IPCC gas reserves (Fig. **S12**) are underestimated, the IPCC case in Fig. (**6**) remains valid if added gas reserves are used at facilities where CO₂ is captured.

However, even with phase-out of coal emissions and assuming IPCC oil and gas reserves, CO₂ would remain above 350 ppm for more than two centuries. Ongoing Arctic and ice sheet changes, examples of rapid paleoclimate change, and other criteria cited above all drive us to consider scenarios that bring CO₂ more rapidly back to 350 ppm or less.

4.4. Policy relevance

Desire to reduce airborne CO₂ raises the question of whether CO₂ could be drawn from the air artificially. There are no large-scale technologies for CO₂ air capture now, but with strong research and development support and industrial-scale pilot projects sustained over decades it may be possible to achieve costs ~\$200/tC [81] or perhaps less [82]. At \$100/tC, the cost of removing 50 ppm of CO₂ is ~\$10 trillion.

Improved agricultural and forestry practices offer a more natural way to draw down CO₂. Deforestation contributed a net emission of 60±30 ppm over the past few hundred years, of

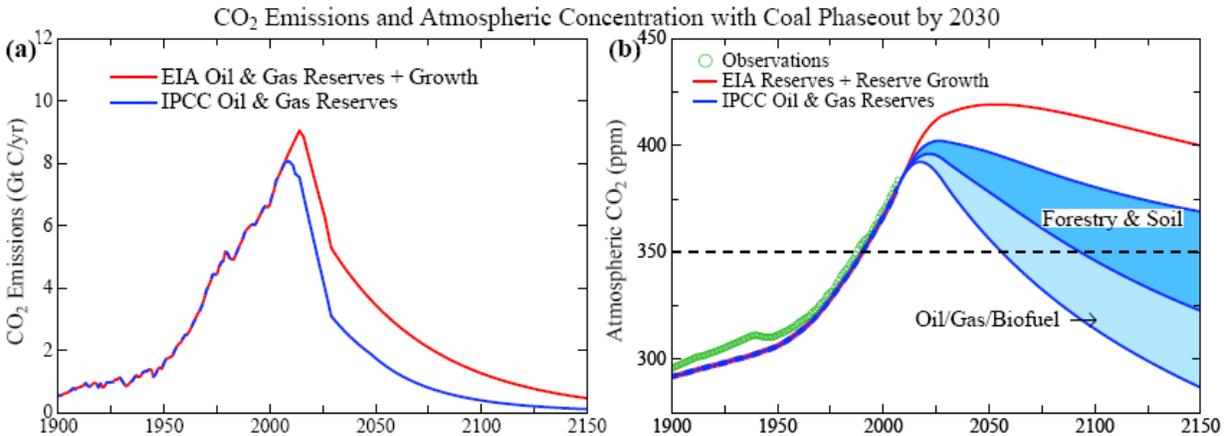


Fig. (6). (a) Fossil fuel CO₂ emissions with coal phase-out by 2030 based on IPCC [2] and EIA [80] estimated fossil fuel reserves. (b) Resulting atmospheric CO₂ based on use of a dynamic-sink pulse response function representation of the Bern carbon cycle model [78, 79].

which ~ 20 ppm CO₂ remains in the air today [2, 83, Figs. (S12), (S14)]. Reforestation could absorb a substantial fraction of the 60 ± 30 ppm net deforestation emission.

Carbon sequestration in soil also has significant potential. Biochar, produced in pyrolysis of residues from crops, forestry, and animal wastes, can be used to restore soil fertility while storing carbon for centuries to millennia [84]. Biochar helps soil retain nutrients and fertilizers, reducing emissions of GHGs such as N₂O [85]. Replacing slash-and-burn agriculture with slash-and-char and use of agricultural and forestry wastes for biochar production could provide a CO₂ drawdown of ~ 8 ppm or more in half a century [85].

In the Appendix we define a forest/soil drawdown scenario that reaches 50 ppm by 2150 (Fig. 6b). This scenario returns CO₂ below 350 ppm late this century, after about 100 years above that level.

More rapid drawdown could be provided by CO₂ capture at power plants fueled by gas and biofuels [86]. Low-input high-diversity biofuels grown on degraded or marginal lands, with associated biochar production, could accelerate CO₂ drawdown, but the nature of a biofuel approach must be carefully designed [85, 87-89].

A rising price on carbon emissions and payment for carbon sequestration is surely needed to make drawdown of airborne CO₂ a reality. A 50 ppm drawdown via agricultural and forestry practices seems plausible. But if most of the CO₂ in coal is put into the air, no such “natural” drawdown of CO₂ to 350 ppm is feasible. Indeed, if the world continues on a business-as-usual path for even another decade without initiating phase-out of unconstrained coal use, prospects for avoiding a dangerously large, extended overshoot of the 350 ppm level will be dim.

5. SUMMARY

Humanity today, collectively, must face the uncomfortable fact that industrial civilization itself has become the principal driver of global climate. If we stay our present course, using fossil fuels to feed a growing appetite for energy-intensive life styles, we will soon leave the climate of the Holocene, the world of prior human history. The eventual response to doubling pre-industrial atmospheric CO₂ likely would be a nearly ice-free planet, preceded by a period of chaotic change with continually changing shorelines.

Humanity's task of moderating human-caused global climate change is urgent. Ocean and ice sheet inertias provide a buffer delaying full response by centuries, but there is a danger that human-made forcings could drive the climate system beyond tipping points such that change proceeds out of our control. The time available to reduce the human-made forcing is uncertain, because models of the global system and critical components such as ice sheets are inadequate. However, climate response time is surely less than the atmospheric lifetime of the human-caused perturbation of CO₂. Thus remaining fossil fuel reserves should not be exploited without a plan for retrieval and disposal of resulting atmospheric CO₂.

Paleoclimate evidence and ongoing global changes imply that today's CO₂, about 385 ppm, is already too high to maintain the climate to which humanity, wildlife, and the rest of the biosphere are adapted. Realization that we must reduce the current CO₂ amount has a bright side: effects that had begun to seem inevitable, including impacts of ocean acidification, loss of fresh water supplies, and shifting of climatic zones, may be averted by the necessity of finding an energy course beyond fossil fuels sooner than would otherwise have occurred.

We suggest an initial objective of reducing atmospheric CO₂ to 350 ppm, with the target to be adjusted as scientific understanding and empirical evidence of climate effects accumulate. Although a case already could be made that the eventual target may need to be lower, the 350 ppm target is sufficient to qualitatively change the discussion and drive fundamental changes in energy policy. Limited opportunities for reduction of non-CO₂ human-caused forcings are important to pursue but do not alter the initial 350 ppm CO₂ target. This target must be pursued on a timescale of decades, as paleoclimate and ongoing changes, and the ocean response time, suggest that it would be foolhardy to allow CO₂ to stay in the dangerous zone for centuries.

A practical global strategy almost surely requires a rising global price on CO₂ emissions and phase-out of coal use except for cases where the CO₂ is captured and sequestered. The carbon price should eliminate use of unconventional fossil fuels, unless, as is unlikely, the CO₂ can be captured. A reward system for improved agricultural and forestry practices that sequester carbon could remove the current CO₂ overshoot. With simultaneous policies to reduce non-CO₂ greenhouse gases, it appears still feasible to avert catastrophic climate change.

Present policies, with continued construction of coal-fired power plants without CO₂ capture, suggest that decision-makers do not appreciate the gravity of the situation. We must begin to move now toward the era beyond fossil fuels. Continued growth of greenhouse gas emissions, for just another decade, practically eliminates the possibility of near-term return of atmospheric composition beneath the tipping level for catastrophic effects.

The most difficult task, phase-out over the next 20-25 years of coal use that does not capture CO₂, is Herculean, yet feasible when compared with the efforts that went into World War II. The stakes, for all life on the planet, surpass those of any previous crisis. The greatest danger is continued ignorance and denial, which could make tragic consequences unavoidable.

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