Target Atmospheric CO₂: Where Should Humanity Aim?

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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 425±75 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.

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 ¾°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 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.

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Stabilizing atmospheric CO$_2$ and climate requires that net CO$_2$ emissions approach zero (10), because of the long lifetime of CO$_2$.

We use paleoclimate data to show that long-term climate has high sensitivity to climate forcings and that the present global mean CO$_2$, 385 ppm, is already in the dangerous zone. Despite rapid current CO$_2$ growth, ~2 ppm/year, we show that it is conceivable to lower CO$_2$ this century to less than the current amount, but only via prompt policy changes.

**Climate sensitivity.** A global climate forcing, measured in W/m$^2$ averaged over the planet, is an imposed perturbation of the planet’s energy balance. Increase of solar irradiance (So) by 2% and doubling of atmospheric CO$_2$ are each forcings of about 4 W/m$^2$ (11).

Charney (12) defined an idealized climate sensitivity problem, asking how much global surface temperature would increase if atmospheric CO$_2$ 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 CO$_2$ 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 (12) used climate models to estimate fast-feedback doubled CO$_2$ 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.

**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/m$^2$, as shown by considering the contrary: an imbalance of 1 W/m$^2$ 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 slow-feedback ice age surface properties, i.e., increased ice area, different vegetation distribution, and continental shelf exposure, was -3.5 ± 1 W/m$^2$ (13) relative to the Holocene. Additional forcing due to reduced amounts of long-lived GHGs (CO$_2$, CH$_4$, N$_2$O), including the indirect effects of CH$_4$ on tropospheric ozone and stratospheric water vapor (fig. S1) was -3 ± 0.5 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 ± 1°C relative to the Holocene (14, 15) yield an empirical sensitivity ~¾ ± ¼ °C per W/m$^2$ forcing, i.e., a Charney sensitivity of 3 ± 1 °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 (11). 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.

**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 CO₂ and CH₄ data from the Antarctic Vostok ice core (16) and sea level based on Red Sea sediment cores (17). 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.

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

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Fe\text{(GHGs)} = 1.12 \left[ Fa\text{(CO}_2\text{)} + 1.4 Fa\text{(CH}_4\text{)} \right],
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using published formulae for Fa of each gas (18). 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 (11). 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 ¾°C per W/m² yields the blue curve in Fig. 1(C). Vostok temperature change (16) divided by two (red curve) is used to crudely estimate global temperature change, as typical glacial-interglacial global annual-mean temperature change is ~5°C and is associated with ~10°C change on Antarctica (19). Figure 1C shows that fast-feedback climate sensitivity ¾°C per W/m² (3°C for doubled CO₂) is a good approximation for the entire period.

**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 (20). 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 large climate swings. Instead changes of GHGs and sea level (a measure of ice sheet size) lag temperature change by typically several hundred years (6, 7, 21, 22).

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 (23, 24). 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 (25).

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, 26). Surface albedo feedback increases as Earth becomes colder and the area of ice increases. Climate sensitivity on 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, may become important on longer geologic time scales.

Paleoclimate data permit evaluation of long-term sensitivity to specified GHG change. We only assume that the area of ice is a function of global temperature. Plotting GHG forcing (7) from ice core data (27) 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. GHG amounts will decline if emissions decrease enough, but, on the other hand, if the globe warms much further, carbon cycle models (2) and empirical data (6, 26) find a positive GHG feedback. Amplification of GHGs is moderate if warming is kept within the range of recent interglacial periods (6), but larger warming risks greater release of CH₄ and CO₂ from methane hydrates in tundra and ocean sediments (28).
**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 (29), as discussed in the Supplementary material.

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 (30-32) reveal frequent ‘suborbital’ sea level changes at rates of 1 m/century or more.

Present-day observations of Greenland and Antarctica show increasing surface melt (33), loss of buttressing ice shelves (34), accelerating ice streams (35), and increasing overall mass loss (36). These rapid changes do not occur in existing ice sheet models, which are missing critical physics of ice sheet disintegration (37). Sea level changes of several meters per century occur in the paleoclimate record (30, 31), 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.

**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 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 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.

**Cenozoic Era.**

Pleistocene atmospheric CO₂ variations occur as a climate feedback, with carbon exchanged among the ocean, atmosphere, soils and biosphere. On longer time scales CO₂ is exchanged between these surface reservoirs and the solid earth, making CO₂ a primary agent of long-term climate change and orbital effects a ‘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 (24) provides a starting point for analyzing Cenozoic climate change (Fig. 3A). At times with negligible ice sheets, oxygen isotope change, δ¹⁸O, provides a direct measure of deep ocean temperature (Tdo). Thus Tdo (°C) ~ -4 δ¹⁸O + 12 between 65.5 and 34 My BP.

Rapid increase of δ¹⁸O at about 34 My is associated with glaciation of Antarctica (24, 38) and global cooling, as evidenced by data from North America (39) and Asia (40). From then until the present, ¹⁸O in deep ocean foraminifera is affected by both ice volume and Tdo, lighter ¹⁶O evaporating preferentially from the ocean and accumulating in ice sheets. Between 34 My and the last ice age (20 ky) the change of δ¹⁸O was ~ 3‰, change of Tdo was ~ 6°C (from +5 to -1°C) and ice volume change ~ 180 msl (meters of sea level). Given that a 1.5‰ change of δ¹⁸O is associated with a 6°C Tdo change, we assign the remaining δ¹⁸O change to ice volume linearly at the rate 60 msl per mil δ¹⁸O change (thus 180 msl for δ¹⁸O between 1.75 and 4.75). Equal division of δ¹⁸O 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 δ¹⁸O yields deep ocean temperature Tdo (°C) = -2 (δ¹⁸O -4.25‰) after 34 My, as in Fig. 3B.

The large (~14°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 ~0.4% in the Cenozoic (41), a linear forcing change of only +1 W/m² and of the wrong sign to contribute to the cooling trend. Climate forcing due to continental locations was < 1 W/m², 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.

CO₂ concentration, in contrast, varied from ~180 ppm in glacial times to 1500 ± 500 ppm in the early Cenozoic (42). This change is a forcing of more than 10 W/m² (Table 1 in 15), an order of magnitude larger than other known forcings. CH₄ and N₂O, 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 (43, 44).

**Cenozoic carbon cycle.** Solid Earth sources and sinks of CO₂ are generally not balanced (45). 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 (46). CO2 returns primarily via metamorphism and volcanic outgassing where carbonate-rich oceanic crust is subducted beneath moving continental plates.

Burial and outgassing of CO2 are each typically 2-4×10^{12} mol C/year (46, 47). An imbalance of 2×10^{12} mol C/year, confined to the atmosphere, is ~0.01 ppm/year, but as CO2 is distributed among surface reservoirs, it is only ~0.0001 ppm/year. This is a negligible rate compared to the present human-made atmospheric CO2 increase of ~2 ppm/year, yet over a million years this small crustal imbalance alters atmospheric CO2 by 100 ppm.

Between 60 and 50 My ago India moved north rapidly, 18-20 cm/year (48), through a region that long had been a depocenter for carbonate and organic sediments. Subduction of carbonate-rich crust was surely a source of CO2 outgassing and a prime cause of global warming, which peaked 50 My ago (Fig. 3b) with the Indo-Asian collision that initiated uplift of the Himalayas and Tibetan Plateau that subsequently increased drawdown of atmospheric CO2 by enhanced weathering (49). Since then the Indian and Atlantic Oceans have been the major depocenters for carbon, with subduction of carbon-rich crust limited largely to small regions near Indonesia and Central America (46).

Thus atmospheric CO2 declined in the past 50 My (42) and climate cooled (Fig. 3B) leading to Antarctic glaciation by ~34 My. Antarctica has remained glaciated ever since, although glaciation may have reversed temporarily, e.g., ~26 My ago, perhaps due to negative feedback of reduced weathering (50).

Knowledge of Cenozoic CO2 is limited to imprecise proxy measures except for recent ice core data. However, the proxy data indicate that CO2 was ~1000-2000 ppm in the early Cenozoic and <500 ppm in the last 20 My (2, 42).

**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 0.3°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 δ^{18}O, we obtain the GHG forcing history Fig. 4C).
We hinge our calculations 35 My ago 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 CO2 amount that precipitated Antarctic glaciation. Finally, the relation between global surface air temperature change ($\Delta T_s$) and deep ocean temperature change ($\Delta T_{do}$) differs for ice-free and glaciated worlds.

In the ice-free world (65-35 My) we take $\Delta T_s \sim \Delta T_{do}$ with generous (50%) uncertainty, as climate models show that global temperature change is tied closely to ocean temperature change (51). In the glaciated world $\Delta T_{do}$ is limited by the freezing point in the deep ocean. $\Delta T_s$ between the last ice age (20 ky) and the present interglacial period (~5°C) was ~1.5 times larger than $\Delta T_{do}$. In fig. S5 we show that this relationship fits well throughout the period of ice core data.

If we specify CO2 at 35 My, the GHG forcing defines CO2 at other times, assuming CO2 provides 75% of the GHG forcing, as in the late Pleistocene. CO2 ~400-450 ppm at 35 My keeps CO2 in the range of early Cenozoic proxies (Fig. 5A) and yields a good fit to the amplitude and mean CO2 amount in the late Pleistocene (Fig. 5B). A ~500 ppm CO2 threshold for Antarctic glaciation was previously inferred from proxy CO2 data and a carbon cycle model (52).

Individual CO2 proxies (Fig. S10) clarify limitations due to scatter among the measurements. Low CO2 of some early Cenozoic proxies suggests higher climate sensitivity. However, in general the sensitivities inferred from the Cenozoic and Phanerozoic (53, 54, 55) 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).

A CO2 estimate of ~425 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 $\Delta T_s$ to $\Delta T_{do}$. The model fails to account for cooling in the past 15 My if CO2 increased, as most 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.
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; 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).

on temperature (56). 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.

**Implications.** We infer from Cenozoic data that CO₂ was the dominant Cenozoic forcing, that CO₂ was ~350-500 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 half way between points A and C in global temperature (Fig. 3B) and climate forcings (Fig. 4). For example, the climate forcing for CO₂ change from 180 to 285 ppm is 2.6 W/m² and further change from 285 to 450 ppm is 2.7 W/m².

Thus equilibrium climate sensitivity that includes slow feedbacks is about 1.5°C per W/m² or 6°C for doubled CO₂ between today and an ice-free world, the same as between today and the last ice age. Evidently amplification provided by loss of Greenland and Antarctic ice and spread of flora over the vast high-latitude land area in the Northern Hemisphere, in response to positive forcing, is comparable to amplification provided by the Laurentide and other ice sheets, in response to negative forcing.
Anthropocene Era.

Human-made global climate forcings now prevail over natural forcings (Fig. 2). Earth may have entered the Anthropocene era (57, 58) 6-8 ky ago (59), 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).

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, 60). 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 (61). West Antarctic ice loss can be accelerated by several feedbacks, once ice loss is substantial (37).

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 (62), ice streams (63), ice shelves (64), and expansion of the subtropics (65, 66).

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.

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 GHG climate forcing change is determined mainly by CO2 (67). 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 (18).

We define a target CO2 level by considering several specific climate impacts:

Civilization is adapted to climate zones of the Holocene. Theory and models indicate that subtropical regions expand poleward with global warming (2, 65). Data reveal a 4-degree latitudinal shift already (66), larger than model predictions, yielding increased aridity in southern
United States, the Mediterranean region, Australia and parts of Africa. Impacts of this climate shift (69) support the conclusion that 385 ppm CO₂ is already deleterious.

Alpine glaciers are in near-global retreat (69, 70). After a flush of fresh water, glacier loss foretells long summers 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 (31, 17, 30). Accelerating mass losses from Greenland (71) and West Antarctica (72) heighten concerns about ice sheet stability. An initial CO₂ target of 350 ppm, to be reassessed as the effect on ice sheet mass balance is 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 (73). 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 (74). 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.

CO₂ scenarios. A large fraction of fossil fuel CO₂ emissions stays in the air a long time, one-quarter remaining airborne for several centuries (5, 75-77). 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, 77). 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 (78) 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.

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 (79) or perhaps less (80). 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 which ~20 ppm CO₂ remains in the air today (2, 81, figs S12, S14). Reforestation could absorb a significant 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 (82). Biochar helps soil retain nutrients and fertilizers, reducing emissions of GHGs such as N₂O (83). 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 in half a century (83).

In Supplementary Material 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 (84). 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 (83, 85-87).

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.

**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.

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 CO$_2$. Thus remaining fossil fuel reserves should not be exploited without a plan for retrieval and disposal of resulting atmospheric CO$_2$.

Paleoclimate evidence and ongoing global changes imply that today’s CO$_2$, 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 CO$_2$ 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 CO$_2$ to 350 ppm, with the target to be adjusted as scientific understanding and empirical evidence of climate effects accumulate. Limited opportunities for reduction of non-CO$_2$ human-caused forcings are important to pursue but do not alter the initial 350 ppm CO$_2$ 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 CO$_2$ to stay in the dangerous zone for centuries.

A practical global strategy almost surely requires a rising global price on CO$_2$ emissions and phase-out of coal use except for cases where the CO$_2$ is captured and sequestered. The carbon price should eliminate use of unconventional fossil fuels, unless, as is unlikely, the CO$_2$ can be captured. A reward system for improved agricultural and forestry practices that sequester carbon could remove the current CO$_2$ overshoot. With simultaneous policies to reduce non-CO$_2$ greenhouse gases, it appears still feasible to avert catastrophic climate change.

Present policies, with continued construction of coal-fired power plants without CO$_2$ 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 CO$_2$, 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.

References and Notes
69. Intergovernmental Panel on Climate Change (IPCC), *Impacts, Adaptation and Vulnerability*, M. Parry et al., Eds. (Cambridge Univ. Press, New York, 2007).
88. We thank H. Harvey, G. Lenfest, the Rockefeller Foundation, and NASA program managers D. Anderson and J. Kaye for research support, S. Baum, P. Essunger, K. Farnish, Q. Fu, L.D. Harvey, I. Horovitz, C. Kutscher, J. Leventhal, C. McGrath, T. Noerpel, P. Read, J. Romm, D. Sanborn, S. Schwartz, K. Ward and S. Weart for comments on a draft manuscript, and NOAA Earth System Research Laboratory for data.