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, the planet being nearly ice-free until CO₂ fell to 450 ± 100 ppm; barring prompt policy changes, the critical level will be passed, in the opposite direction, within decades. 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, but likely less than that. 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 (re-lative to 2000, 1.7°C re-lative to pre-industrial time), aiming to avoid practically irreversible ice sheet and species loss. This 1°C limit, with nominal climate sensitivity of ¼°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 ‘low’ 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 be significant contributors to climate change. 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 a 1°C atmospheric CO₂ and climate requires that net CO₂ emissions approach zero, because of the long lifetime of CO₂ [10, 11].

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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 (So) by 2% and doubling of atmospheric CO₂ are each forcing 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, are 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 means in terms of fast and slow feedbacks is useful, even though time scales of these changes may vary. We include as fast feedbacks aerosol changes, e.g., of desert dust and marine dimethylsulfide, that occur in response 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.

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Climate forcing in the LGM equilibrium state due to the ice age surface properties, i.e., increased ice sheet area, different vegetation distribution, and continental shelf exposure, was ~3.5 ± 1 W/m² [14] relative to the Holocene. Additional forcing due to reduced amounts of long-lived GHGs (CO₂, CH₄, N₂O), including the indirect effects of CH₄ on tropospheric ozone and stratospheric water vapor (Fig. S1) was ~3 ± 0.5 W/m². Global forcing due to slight changes in the Earth's orbit is a negligible fraction of 1 W/m² (Fig. S3). The total 6.5 W/m² forcing and global surface temperature change of 5 ± 1°C relative to the Holocene [15, 16] yield an empirical sensitivity of 3 ± 1°C per W/m² forcing, i.e., a Charney sensitivity of 3 ± 1 °C for the 4 W/m² forcing of doubled CO₂. 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 greenhouse effect [12]. At its present temperature Earth is on a flat portion of its fast-feedback climate sensitivity curve (Fig. S2). Thus our empirical sensitivity, 1.5°C per W/m², is consistent with the Earth's current flat portion of fast-feedback climate sensitivity curve (Fig. S2).

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]. Second, we take the LGM and late Holocene as 3 W/m², apportioned as 75% CO₂, 14% CH₄ and 11% N₂O. 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.

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2. PLEISTOCENE EPOCH

Atmospheric composition and surface properties in the late Pleistocene are known well enough for accurate assessment of fast-feedback (Charney) climate sensitivity. We first compare the pre-industrial Holocene with the last glacial maximum [LGM, 20 ky B.P. (before present)]. The Earth's history, however, allows empirical inference of both fast feedback climate sensitivity and long-term sensitivity to specified GHG change including t he low ice sheet area, and high temperatures at high latitudes.

Climate forcing in the LGM equilibrium state due to the ice age surface properties, i.e., increased ice sheet area, different vegetation distribution, and continental shelf exposure, was ~3.5 ± 1 W/m² [14] relative to the Holocene. Additional forcing due to reduced amounts of long-lived GHGs (CO₂, CH₄, N₂O), including the indirect effects of CH₄ on tropospheric ozone and stratospheric water vapor (Fig. S1) was ~3 ± 0.5 W/m². Global forcing due to slight changes in the Earth's orbit is a negligible fraction of 1 W/m² (Fig. S3). The total 6.5 W/m² forcing and global surface temperature change of 5 ± 1°C relative to the Holocene [15, 16] yield an empirical sensitivity of 3 ± 1 °C per W/m² forcing, i.e., a Charney sensitivity of 3 ± 1 °C for the 4 W/m² forcing of doubled CO₂. 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.

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Fig. (1). (a) CO₂, CH₄ [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 3.3°C per W/m². Observations are Antarctic temperature change [18] divided by two.

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. (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 ~5 °C and is associated with ~1 °C change on Antarctica [21]. Fig. (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.

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 polar ice cores and sea surface temperature derived from ocean sediment 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 sheets 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]. A slow-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 CO$_2$ 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 Pleistocene timescales includes slow feedbacks, and is larger than the Charney sensitivity, because the dominant slow feedbacks are positive. Other feedbacks, e.g., the negative feedback of increased weathering as CO$_2$ increases, become important on longer geologic time scales.

Paleoclimate data permit evaluation of long-term sensitivity to specified GHG change. We assume only that, to first order, 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/m$^2$ or 6°C for doubled CO$_2$ (Fig. 2), twice as large as the Charney feedback sensitivity. 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 CH$_4$ and CO$_2$ from methane hydrates in tundra and ocean sediments [29]. On still longer, geological, timescales weathering of rocks causes a negative feedback on atmospheric CO$_2$ 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 with specified GHG change? Response is slowed by ocean thermal inertia and the time needed for ice sheets to disintegrate.

Ocean-caused de lay 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 Supplementary Material Section.

Ice sheet response time is often assumed to be several millennia, but sed on the broad scale 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 but tressing ice shelves [36], acelerating ice streams [37], and d in creasing 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 glacial te mperature (purple curve) and an equilibrium (long-term) temperature response based on the net estimated climate forcing (black 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). At climate model width includes only fast feedbacks, has additional warming of ~0.6°C in the current atmospheric composition and actual global temperature [26-33]. Subtracting the ice volume change of ~1.4°C the effective feedback is increase of atmospheric CO2 as climate changes. The forcing in Fig. 2 (black curve) reveals a growing gap between actual glacial temperature (purple curve) and an 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 remaining gap between equilibrium temperature for current atmospheric composition and actual global temperature is ~1.4°C. This further 1.4°C warming might be due to ice sheet melt and surface warming of ~0.6°C in the pipeline today because of ocean thermal inertia [5, 8].

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3. CENOZOIC ERA

Pleistocene atmospheric CO2 variations occur as a climate feedback, as carbon is exchanged among surface reservoirs: the ocean, atmosphere, soils and biosphere. The most effective feedback is increase of atmospheric CO2 as climate warm, t he CO2 tr ansfers to the oceans. In contrast, CO2 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/m2 (Table 1 in [16]), an order of magnitude larger than other known forcings. C H4 and N2O, positively correlated with CO2 in the global temperature record, are 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 CO$_2$ are not, in general, balanced at any given time [30, 47]. CO$_2$ 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]. CO$_2$ returns primarily via metamorphism and volcanic outgassing at locations where carbonate-rich oceanic crust is being subducted beneath moving continental plates.

Outgassing and burial of CO$_2$ are each typically $10^{12-13}$ 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 $\sim 10^{12}$ mol C/year cannot be maintained over thousands of years. Such an imbalance, if confined to the atmosphere, would be $\sim 0.005$ ppm/year, but as CO$_2$ is distributed among surface reservoirs, its size on a $\sim 0.0001$ ppm/year. This rate is negligible compared to the present human-made atmospheric CO$_2$ increase of $\sim 2$ ppm/year, yet over a million years such a crustal imbalance alters atmospheric CO$_2$ by 100 ppm.

Between 60 and 50 My ago India moved north rapidly, 18-20 cm/year [50], through a region that had been a depocenter for carbonate and organic sediments. Subduction of carbon-rich crust was surely a large source of CO$_2$ outgassing and a prime cause of global warming, which peaked 50 My ago (Fig. 3b) with the Indo-Asian collision. CO$_2$ must have then decreased due to a reduced subduction source and enhanced weathering with uplift of the Himalayas/Tibetan Plateau [51]. Since then, the Indian and Atlantic Oceans have been major depocenters for carbon, but subduction of carbon-rich crust has been limited mainly to small regions near Indonesia and Central America [47].

Thus atmospheric CO$_2$ declined following the Indo-Asian collision [44] and climate cooled (Fig. 3b) leading to Antarctic glaciation by $\sim 34$ My. Antarctica has been more or less glaciated ever since. The rate of CO$_2$ drawdown declines as atmospheric CO$_2$ 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 CO$_2$, which has been equal within 1-2 percent over the past 700 ky [52]. Large fluctuations in the size of the Antarctic ice sheet have occurred in the past 34 My, possibly related to temporal variations of plate tectonics [53] and outgassing rates. The relatively constant a m
pheric CO$_2$ amount of the past 20 My (Fig. S10) implies a near balance of outgassing and weathering rates over that period.

Knowledge of Cenozoic CO$_2$ 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 CO$_2$ 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 CO$_2$

The entire Cenozoic climate forcing history (Fig. 4a) is implied by the temperature reconstruction (Fig. 3b), assuming a fast-feedback sensitivity of $\frac{2}{3}$°C per W/m$^2$. Subtracting the solar and surface albedo forcings (Fig. 4b), the latter from Eq. S2 with ice sheet area vs time from $\delta^{18}$O, we obtain the GHG forcing history (Fig. 4c).

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 ($\Delta T_s$) and deep ocean temperature change ($\Delta T_{do}$) 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 $\Delta T_{do}$ is limited by the freezing point in the deep ocean. $\Delta T_s$ between the last ice age (20 ky) and the present

![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 $\frac{2}{3}$°C per W/m$^2$. Ratio of $T_s$ change and $T_{do}$ change is assumed to be 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.](P-0003485)
The interglacial period (~5°C) was ~1.5 times larger than $\Delta T_{de}$. 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. $CO_2 \sim 450$ 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 $CO_2$ threshold for Antarctic glaciation of ~500 ppm was previously inferred from proxy $CO_2$ data and a carbon cycle model [5].

Individual $CO_2$ proxies (Fig. S10) clarify limitations due to scatter among the measurements. Low $CO_2$ 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] are in good agreement, 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_e to ΔT_0. The model fails to account for cooling in the past 15 My if CO₂ increased, as proposed by proxies suggesting (Fig. S10). Cing 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. A 4 W/m² forcing carries the planet to equilibrium, to the ice-free state. Thus equilibrium climate sensitivity to GHG change, including 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 underestimate of expected equilibrium global warming in response to imposed doubled CO₂. Additional warming, due to slow climate feedbacks in cluing loose of ice and carbon spread of flora over the vast high-latitude land a rea in the Northern Hemisphere, is approximately double as equilibrium climate sensitivity.

Equilibrium sensitivity 6°C for doubled CO₂ is relevant to the case in which GHG changes are specified. That is appropriate to the anthropogenic process, providing the GHG amounts are estimated from carbon cycle models including climate feedbacks such as methane release from tundra and ocean sediments. The equilibrium sensitivity is even higher if the GHG forcing is included as part of the climate response, as is appropriate for analysis of the climate response to Earth orbital perturbations. The very high sensitivity with both albedo and GHG slow feedbacks included accounts for the huge magnitude of glacial-interglacial fluctuations in the Pleistocene (Fig. 3) in response to small forcings (section 3 of Supplementary Material).

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 suggests that positive feedbacks, especially surface albedo changes, can produce rapid global warming, including sea level rise as fast as several meters per century [7]. Thus if humans push the climate system sufficiently far from 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 pre vail over natural forcings (Fig. 2). Earth may have entered the Anthropocene era [60, 61] 6-8 kya 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 climate 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 a level of the Eocene, 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 change occurs at a point where, without additional forcing, rapid changes proceed practically out of our control [2, 7, 63, 64]. 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 [65]. West Antarctic ice loss can be accelerated by several feedbacks, once ice loss is substantial [30].

We define: (1) the tipping level, the global climate forcing that, if 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 it initiates 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 the hanges of sea ice [65], ice sheets [66], ice shelves [36], and expansion of the subtropics [67, 68].

The tipping level is easier to assess, because the paleoclimate quasi-equilibrium response to known climate forcings 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 of or period of tolerable ove rshoot. However, if overshoot is in place for centuries, the thermal per-
turbation will so penetrate the ocean [10] that recovery without dramatic effects, such as ice sheet disintegration, becomes unlikely.

4.2. Target CO₂

Combined, 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. Thus total GHG climate forcing change is now determined mainly by CO₂ [69]. Coincidentally, CO₂ forcing is similar to the net human-made forcing, because 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 efforts to reduce non-CO₂ GHGs could alleviate the CO₂ requirement, allowing up to about +25 ppm CO₂ for the same climate effect, while re-surgent growth of non-CO₂ GHG s could reduce a lower CO₂ 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 offset 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 limits the rate of climate change, such a CO₂ level likely would cause the passing of climate tipping points and initiate dynamic planetary energy imbalance, but observations of the entire ocean are needed for quantification. CO₂ at a mount must be reduced to 325-355 ppm to increase outgoing flux 0.5-1 W/m², if other forcings are unchanged. A further imbalance reduction, and thus CO₂ ~300-325 ppm, may be needed to restore sea ice to its area of 25 years ago.

Coral reef 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 a climate resembling that to which humanity is accustomed, the climat e of the Holocene, 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 as [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. S12) Keeps maximum CO₂ close to 400 ppm, depending on oil and gas reserves and reserve growth. IPCC reserves assume that half of readily ex tractable oil and gas as [2, 79]. The only realistic way to sharply curtail CO₂ emissions is to phase out coal use except where CO₂ is captured and sequestered.

Civilization is adapted to climate zones of the Holocene. Theory and models indicate that subtropical regions expand poleward with global warming [2, 67]. DaTa are real a 4-degree latitudinal shift aready [68], larger than model predictions, yielding increased aridity in southern United States [70, 71], the Mediterranean region, Aus tralia 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 one-time added flux of fresh water, the Mediterranean region, Aus tralia and the 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]. A accelerating mass losses from Greenland [74] and West Antarctica [75] lightens 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, re-storat on of planetary energy balance. Climate models driven by known forcings yield a present planetary energy imbalance of +0.5-1 W/m². Observed heat increase in the upper 700 m of the ocean [76] confirms the planetary energy imbalance, but observations of the entire ocean are needed for quantification. CO₂ at a mount must be reduced to 325-355 ppm to increase outgoing flux 0.5-1 W/m², if other forcings are unchanged. A further imbalance reduction, and thus CO₂ ~300-325 ppm, may be needed to restore sea ice to its area of 25 years ago.

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4.4. Policy Relevance

Desire to reduce a birnbe C O₂ ra ises t he que stion 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 $200/tC, the cost of removing 50 ppm of CO2 is ~$20 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, 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 agicultural and forestry wastes for biochar production could provide a CO2 drawdown of ~8 ppm or more in half a century [85].

In the Supplementary Material section we define a forest/soil drawdown scenario that achieves 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 decelerate CO2 drawdown, but the nature of a bi ofuel approach must be carefully designed [85, 87-89].

A rising price on carbon emissions and payment for carbon sequestration is surely needed to make a drawdown of CO2 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 une constrained coal use, prospects for avoiding a dangerously large, extended overshoot of the 350 ppm level will be dim.

4.5. Caveats: Climate Variability, Climate Models, and Uncertainties

Climate has great variability, much of which is unforced and unpredictable [2, 90]. This fact raises a practical issue: what is the chance that climate variations, e.g., a temporary cooling trend, will affect public recognition of climate change, making it difficult to implement mitigation policies? Also what are the greatest uncertainties in the expectation of a continued global warming trend? And what are the impacts of climate model limitations, given the inability of models to realistically simulate many aspects of climate change and climate processes?

The El Nino Southern Oscillation (ENSO) [94] accounts for most low latitude temperature variability and much of the global variability. The global impact of ENSO is coherent from month to month, as shown by the global-ocean-mean SST (Figs. 7b), for which the eNSO is an integral part of the El Nino-Southern Oscillation, minimizing the effect of weather noise. The cool anomaly of 2008 coincides with a low ENSO m inimum and a high f temp erature variability, th e planet’s energy imbalance, an d global climate forcings.

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Decadal time scale variability, such as predicted weakening of the Atlantic overturning circulation [95], could interrupt global warming, as discussed in section 18 of the Supplementary Material. But the impact of regional dynamical effects on global temperature is opposed by the planet’s energy imbalance [96], a product of the climate system’s thermal inertia, which includes the warming of the oceans and the atmosphere.

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].
storage [97]. This energy imbalance makes decadal interruption of global warming, in the absence of a negative climate forcing, improbable [96].

Volcanoes and the sun cause significant negative forcings. However, even if the solar irradiance remained at its value in the current solar minimum, this reduced forcing would be offset by increasing CO₂ within seven years (Supplementary Material section 18). Humankind-made aerosols cause a greater net gative forcing, both directly and through their effects on clouds. The first satellite observations of aerosols and clouds with accuracy sufficient to quantify this forcing are planned to begin in 2009 [98], but most analysts anticipate that human-made aerosols will decrease in the future, rather than increase further.

Climate models have many deficiencies in their abilities to simulate climate change [2]. However, model uncertainties cut both ways: it is at least as likely that models underestimate effects of humankind-made GHGs as overestimate them (Supplementary Material section 18). Model deficiencies in evaluating tipping points, the possibility that rapid changes can occur without additional climate forcing [63, 64], are of special concern. Loss of Arctic sea ice, for example, has proceeded more rapidly than predicted by climate models [99]. There are reasons to expect that other nonlinear problems, such as ice sheet disintegration and the extinction of interdependent species and ecosystems, also have the potential for rapid change [39, 63, 64].

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 on present course, using fossil fuels to feed a growing appetite for energy-intensive lifestyles, 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 humankind-caused global climate change is urgent. Ocean and ice sheet in Earth provides a buffer delaying full response by centuries, but there is a danger that humankind-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, be cause models of the global system and critical components such as ice sheets are inadequate. However, climate response times are 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 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 been seen as 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 impacts accumulates. Although a case for a ready CO$_2$ would be made that the eventual target probably needs to be lower, the 350 ppm target is sufficient to qualitatively change the disbursement and drive fundamental changes in energy policy. 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's 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 C$_2$ emissions and a phasing-out 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 growth of greenhouse gas emissions, for we must begin to move now toward the era beyond fossil fuels. Continued growth of greenhouse gases, it appears still feasible to avert catastrophic climate change.

Present poli cies, 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 gases, for just another decade, practically eliminates the possibility of near-term re-turn of a pre-industrial CO$_2$ composition before the tipping level for catastrophic effects.

The most difficult task, phasing over the next 20-25 years of coal use that does not capture CO$_2$, is Herculean, yet feasible when co-mated with the efforts of the international community. This target must be pursued on a timescale of decades, as paleoclimate and ongoing changes, and the ocean's response time, suggest that it would be foolhardy to allow CO$_2$ to stay in the dangerous zone for centuries.

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