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Target Atmospheric CO_2: Where Should Humanity Aim?

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Abstract: Paleoclimate data show that climate sensitivity is ~3°C for doubled CO_2, including only fast feedback processes. Equilibrium sensitivity, including slower surface albedo feedbacks, is ~6°C for doubled CO_2 for the range of climate states between glacial conditions and ice-free Antarctica. Decreasing CO_2 was the main cause of a cooling trend that began 50 million years ago, the planet being nearly ice-free until CO_2 fell to 450 ± 100 ppm; barring prompt policy changes, that 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 CO_2 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-CO_2 forcings. An initial 350 ppm CO_2 target may be achievable by phasing out coal use except where CO_2 is captured and adopting agricultural and forestry practices that sequester carbon. If the present overshoot of this target CO_2 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 ¾°C per W/m^2 and plausible control of other GHGs [6], implies maximum CO_2 ~ 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 CO_2 limit is in the range 300-500 ppm for 25 percent risk tolerance, depending on climate sensitivity and non-CO_2 forcings. Stabilizing atmospheric CO_2 and climate requires that net CO_2 emissions approach zero, because of the long lifetime of CO_2 [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 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 reduce CO_2 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/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 [12].

Charney [13] 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 [13] 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.

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/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 ice age surface properties, i.e., increased ice area, different vegetation distribution, and continental shelf exposure, was -3.5 ± 1 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 ± 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. **S3**). The total 6.5 W/m^2 forcing and global surface temperature change of 5 ± 1°C relative to the Holocene [15, 16] 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 [12]. At its present temperature Earth is on a flat portion of its fast-feedback climate sensitivity curve (Fig. **S2**). 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.

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 N_2O climate forcing as 12 percent of the sum of the CO_2 and CH_4 forcings, rather than the 15 percent estimated earlier [7] Because N_2O data are not available for the entire record, and its forcing is small and highly correlated with CO_2 and CH_4, we take the GHG effective forcing as

$$Fe\,(GHGs) = 1.12\,[Fa(CO_2) + 1.4\,Fa(CH_4)], \quad (1)$$

using published formulae for Fa of each gas [20]. The factor 1.4 accounts for the higher efficacy of CH_4 relative to CO_2, which is due mainly to the indirect effect of CH_4 on tropospheric ozone and stratospheric water vapor [12]. The resulting GHG forcing between the LGM and late Holocene is 3 W/m^2, apportioned as 75% CO_2, 14% CH_4 and 11% N_2O.

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

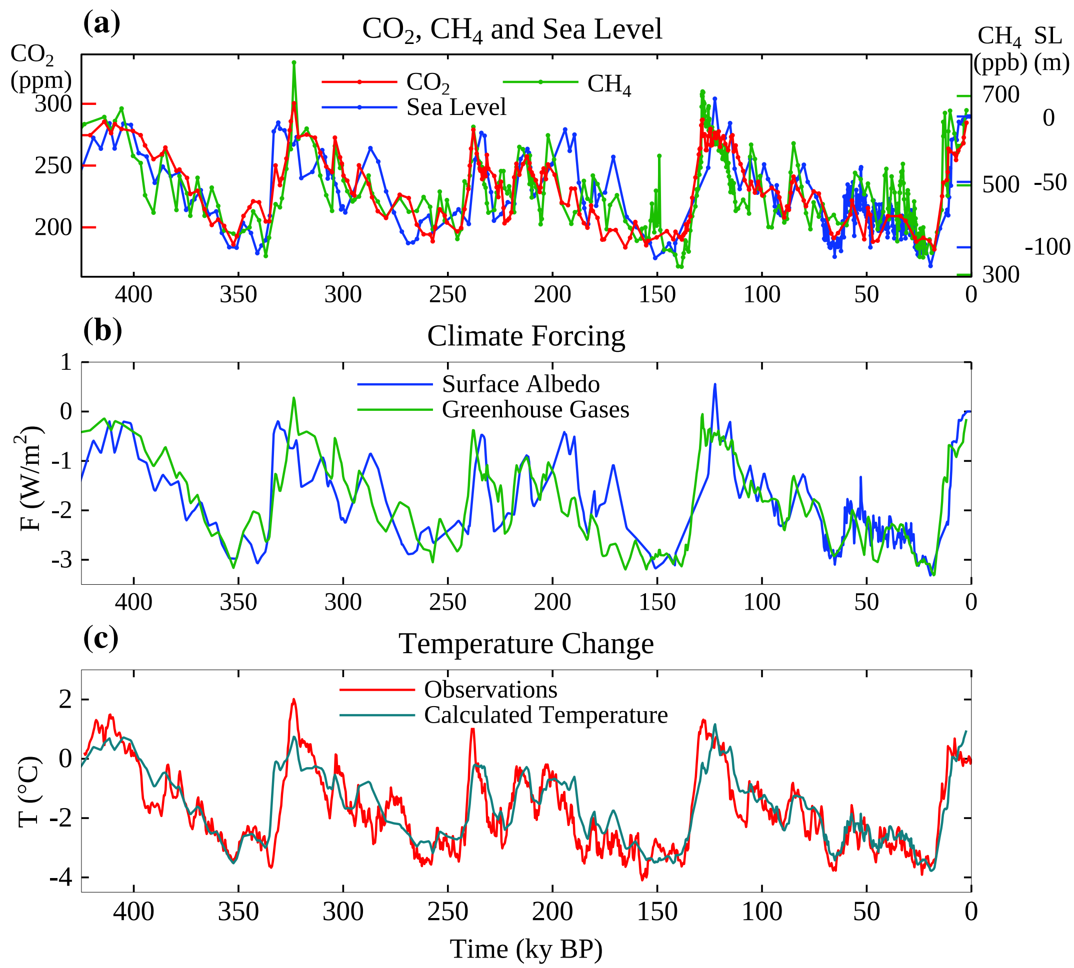

Fig. (1). (**a**) CO_2, CH_4 [17] and sea level [19] for past 425 ky. (**b**) Climate forcings due to changes of GHGs and ice sheet area, the latter inferred from sea level change. (**c**) Calculated global temperature change based on climate sensitivity of ¾°C per W/m^2. Observations are Antarctic temperature 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/m^2 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 ~10°C change on Antarctica [21]. Fig. (**1c**) shows that fast-feedback climate sensitivity ¾°C per W/m^2 (3°C for doubled CO_2) 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 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 CO_2 by the ocean, due partly to temperature dependence of CO_2 solubility but mostly to increased ocean mixing in a warmer climate, which acts to flush out

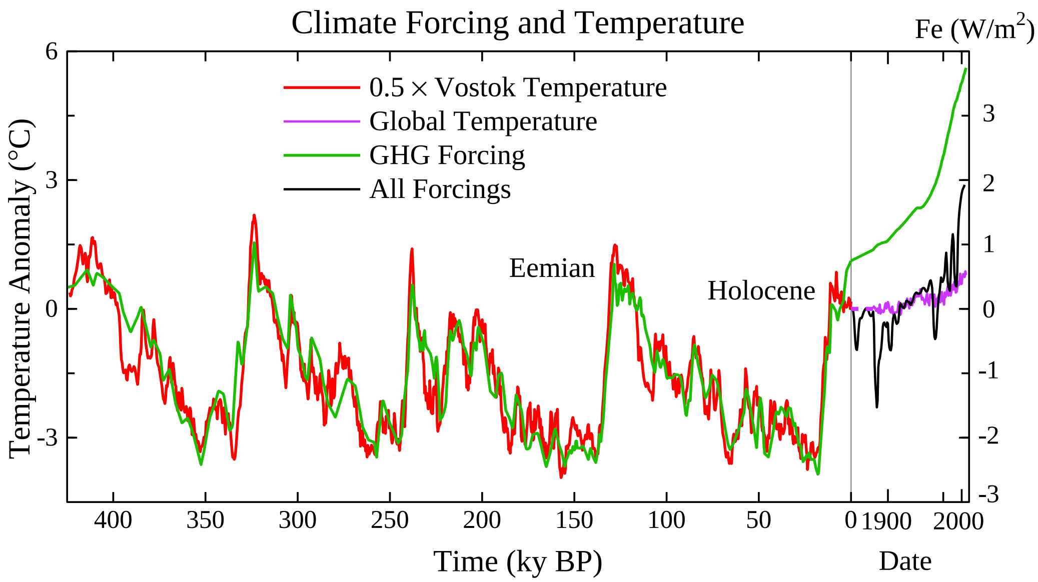

Fig. (2). Global temperature (left scale) and GHG forcing (right scale) due to CO_2, CH_4 and N_2O from the Vostok ice core [17, 18]. Time scale is expanded for the industrial era. Ratio of temperature and forcing scales is 1.5°C per W/m^2, i.e., the temperature scale gives the expected equilibrium response 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.

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 time scales 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 fast-feedback sensitivity. Note that we assume the area of ice and snow on the planet to be predominately dependent on global temperature, but some changes of regional ice sheet properties occur as part of the Earth orbital climate forcing (see Supplementary Material).

This equilibrium sensitivity of 6° C for doubled CO_2 is valid for specified GHG amount, as in studies that employ emission scenarios and coupled carbon cycle/climate models to determine GHG amount. If GHGs are included as a feedback (with say solar irradiance as forcing) sensitivity is still larger on Pleistocene time scales (see Supplementary Material), but the sensitivity may be reduced by negative feedbacks on geologic time scales [29, 30]. The 6°C sensitivity reduces to 3°C when the planet has become warm enough to lose its ice sheets.

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 CH_4 and CO_2 from methane hydrates in tundra and ocean sediments [29]. On still longer, geological, time scales 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 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 Supplementary Material Section.

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 CO_2 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 CO_2 as climate warms, the CO_2 transfer being mainly from ocean to atmosphere [27, 28]. On longer time scales the total amount of CO_2 in the surface reservoirs varies due to exchange of carbon with the solid earth. CO_2 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, $\delta^{18}O$, provides a direct measure of deep ocean temperature (T_{do}). Thus T_{do} (°C) ~ -4 $\delta^{18}O$ + 12 between 65.5 and 35 My BP.

Rapid increase of $\delta^{18}O$ 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 35 My and the last ice age (20 ky) the change of $\delta^{18}O$ was ~ 3‰, change of T_{do} 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 $\delta^{18}O$ is associated with a 6°C T_{do} change, we assign the remaining $\delta^{18}O$ change to ice volume linearly at the rate 60 msl per mil $\delta^{18}O$ change (thus 180 msl for $\delta^{18}O$ between 1.75 and 4.75). Equal division of $\delta^{18}O$ 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 $\delta^{18}O$ yields deep ocean temperature T_{do} (°C) = -2 ($\delta^{18}O$ -4.25‰) after 35 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 [43], a linear forcing change of only +1 W/m^2 and of the wrong sign to contribute to the cooling trend. Climate forcing due to continental locations was < 1 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 CO_2 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 CO_2 [45, 46].

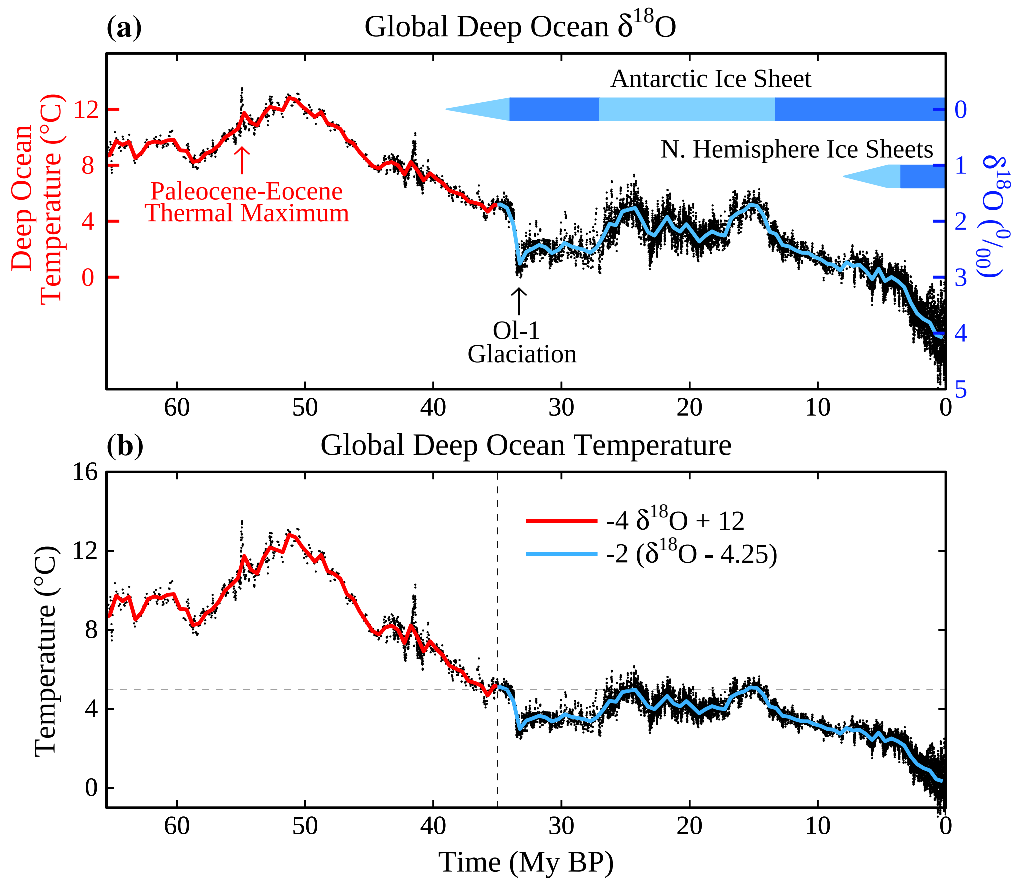

Fig. (3). Global deep ocean (**a**) $\delta^{18}O$ [26] and (**b**) temperature. Black curve is 5-point running mean of $\delta^{18}O$ original temporal resolution, while red and blue curves have 500 ky resolution.

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}-10^{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 ~10^{12} 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 CO_2 is distributed among surface reservoirs, this is only ~ 0.0001 ppm/year. This rate is negligible compared to the present human-made atmospheric CO_2 increase of ~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 long 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 ~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 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 in the past 34 My, possibly related to temporal variations of plate tectonics [53] and outgassing rates. The relatively constant atmospheric 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 ¾°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 (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

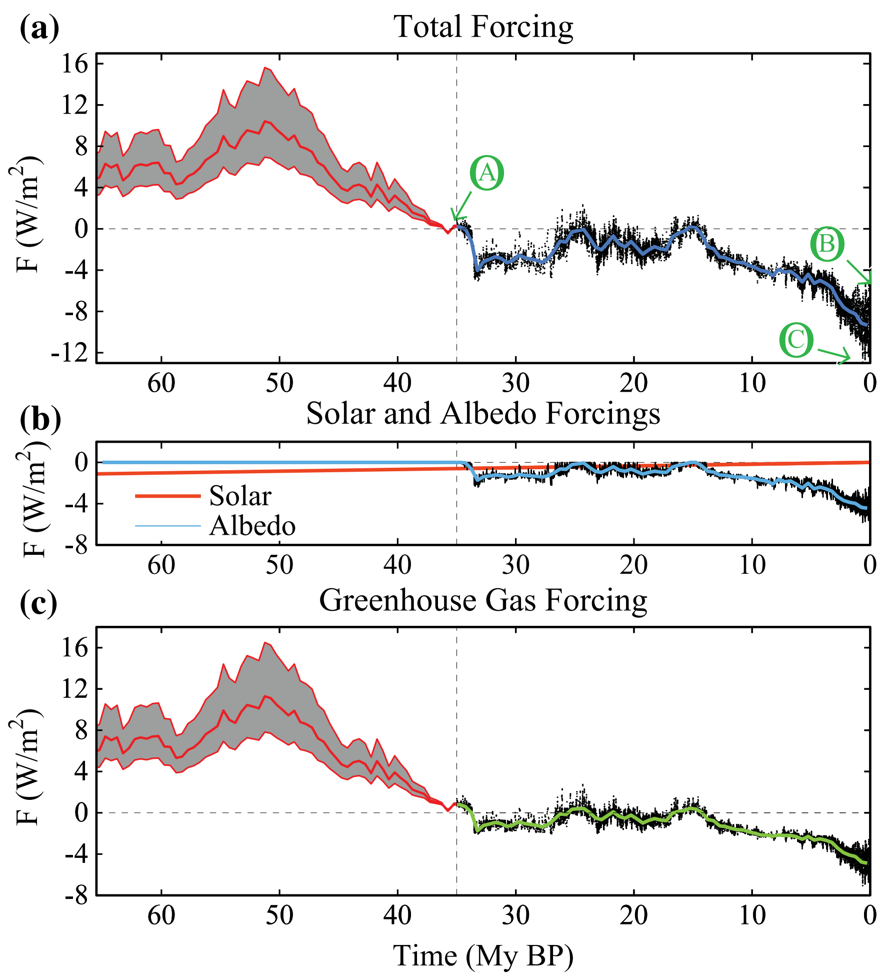

Fig. (4). (**a**) Total climate forcing, (**b**) solar and surface albedo forcings, and (**c**) GHG forcing in the Cenozoic, based on T_{do} history of Fig. (**3b**) and assumed fast-feedback climate sensitivity ¾°C per W/m^2. Ratio of T_s change and T_{do} change is assumed to be near unity in the minimal 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.

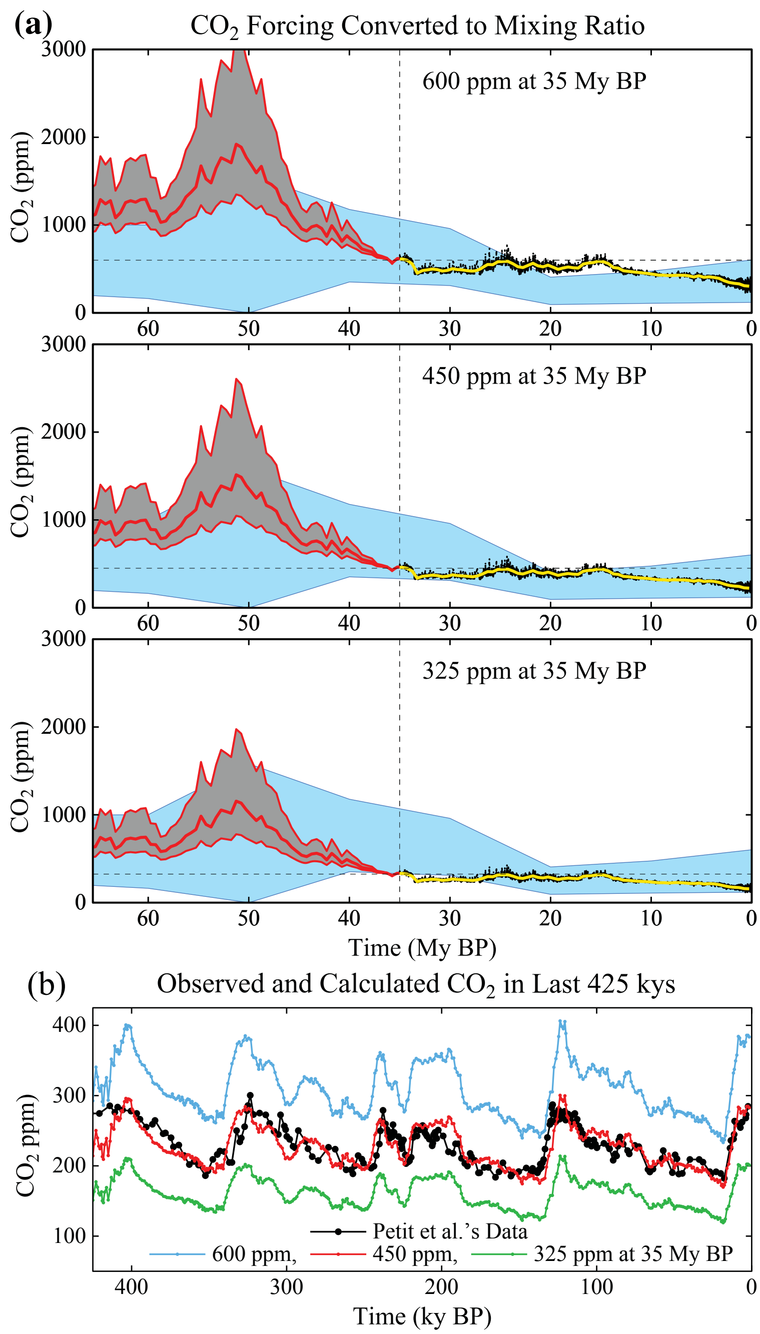

Fig. (5). (**a**) Simulated CO_2 amounts in the Cenozoic for three choices of CO_2 amount at 35 My (temporal resolution of black and colored curves as in Fig. (**3**); blue region: multiple CO_2 proxy data, discussed with Fig. (**S10**); gray region allows 50 percent uncertainty in ratio of global surface and deep ocean temperatures). (**b**) Expanded view of late Pleistocene, including precise ice core CO_2 measurements (black curve).

interglacial period (~5°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. CO_2 ~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 [55].

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] 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 CO_2 estimate of ~450 ppm at 35 My (Fig. **5**) serves as a prediction to compare with new data on CO_2 amount. Model uncertainties (Fig. **S10**) include possible changes of non-CO_2 GHGs and the relation of ΔT_s to ΔT_{do}. The model fails to account for cooling in the past 15 My if CO_2 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-CO_2 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 CO_2 was the dominant Cenozoic forcing, that CO_2 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/m^2. Additional 4 W/m^2 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 CO_2, although valid for the idealized Charney definition of climate sensitivity, is a considerable understatement of expected equilibrium global warming in response to imposed doubled CO_2. 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.

Equilibrium sensitivity 6°C for doubled CO_2 is relevant to the case in which GHG changes are specified. That is appropriate to the anthropogenic case, provided 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 feedback 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 spur rapid global warmings, 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, 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 [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 quasi-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 CO_2

Combined, GHGs other than CO_2 cause climate forcing comparable to that of CO_2 [2, 6], but growth of non-CO_2 GHGs is falling below IPCC [2] scenarios. Thus total GHG climate forcing change is now determined mainly by CO_2 [69]. Coincidentally, CO_2 forcing is similar to the net human-made forcing, because non-CO_2 GHGs tend to offset negative aerosol forcing [2, 5].

Thus we take future CO_2 change as approximating the net human-made forcing change, with two caveats. First, special effort to reduce non-CO_2 GHGs could alleviate the CO_2 requirement, allowing up to about +25 ppm CO_2 for the same climate effect, while resurgent growth of non-CO_2 GHGs could reduce allowed CO_2 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 CO_2 through the Cenozoic Era provides a sobering perspective for assessing an appropriate target for future CO_2 levels. A CO_2 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 CO_2 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 CO_2 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 CO_2 is already deleterious.

Alpine glaciers are in near-global retreat [72, 73]. After a one-time added flush of fresh water, glacier demise will yield 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 CO_2 is already a threat.

Equilibrium sea level rise for today's 385 ppm CO_2 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 CO_2 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/m^2 [5]. 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. CO_2 amount must be reduced to 325-355 ppm to increase outgoing flux 0.5-1 W/m^2, if other forcings are unchanged. A further imbalance reduction, and thus CO_2 ~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 CO_2 is already deleterious. A 300-350 ppm CO_2 target would significantly relieve both of these stresses.

4.3. CO_2 Scenarios

A large fraction of fossil fuel CO_2 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 climate 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 [2, 79]. The only realistic way to sharply curtail CO_2 emissions is to phase out coal use except where CO_2 is captured and sequestered.

Phase-out of coal emissions by 2030 (Fig. **6**) keeps maximum CO_2 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 the additional gas reserves are used at facilities where CO_2 is captured.

However, even with phase-out of coal emissions and assuming IPCC oil and gas reserves, CO_2 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 CO_2 more rapidly back to 350 ppm or less.

4.4. Policy Relevance

Desire to reduce airborne CO_2 raises the question of whether CO_2 could be drawn from the air artificially. There are no large-scale technologies for CO_2 air capture now, but

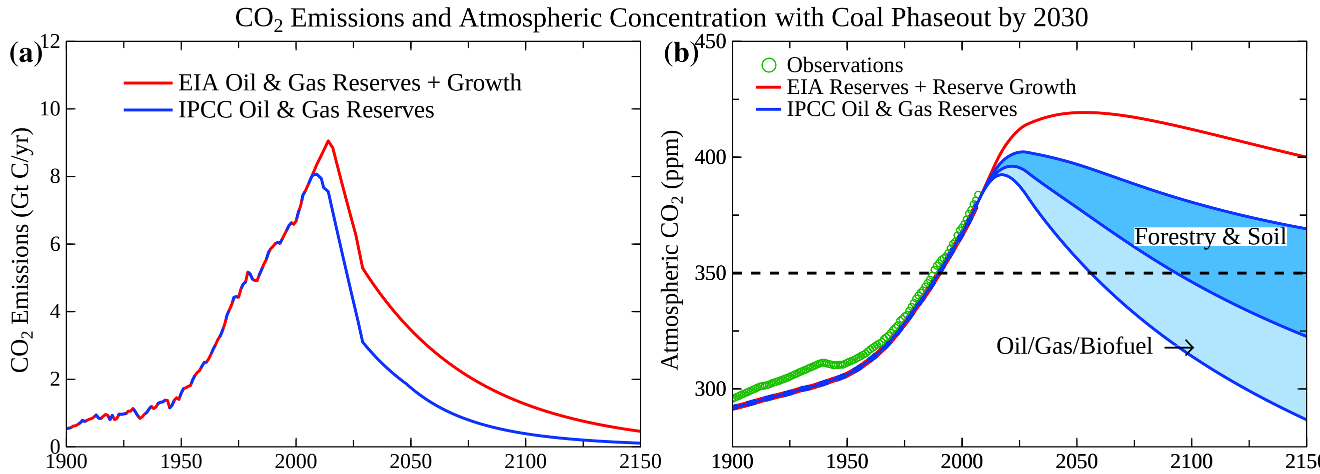

Fig. (6). (**a**) Fossil fuel CO_2 emissions with coal phase-out by 2030 based on IPCC [2] and EIA [80] estimated fossil fuel reserves. (**b**) Resulting atmospheric CO_2 based on use of a dynamic-sink pulse response function representation of the Bern carbon cycle model [78, 79].

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 CO_2 is ~\$20 trillion.

Improved agricultural and forestry practices offer a more natural way to draw down CO_2. Deforestation contributed a net emission of 60±30 ppm over the past few hundred years, of which ~20 ppm CO_2 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 N_2O [85]. Replacing slash-and-burn agriculture with slash-and-char and use of agricultural and forestry wastes for biochar production could provide a CO_2 drawdown of ~8 ppm or more in half a century [85].

In the Supplementary Material Section we define a forest/soil drawdown scenario that reaches 50 ppm by 2150 (Fig. **6b**). This scenario returns CO_2 below 350 ppm late this century, after about 100 years above that level.

More rapid drawdown could be provided by CO_2 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 CO_2 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 CO_2 a reality. A 50 ppm drawdown *via* agricultural and forestry practices seems plausible. But if most of the CO_2 in coal is put into the air, no such "natural" drawdown of CO_2 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.

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 atmosphere and ocean exhibit coupled nonlinear chaotic variability that cascades to all time scales [91]. Variability is so large that the significance of recent decadal global temperature change (Fig. **7a**) would be very limited, if the data were considered simply as a time series, without further information. However, other knowledge includes information on the causes of some of the temperature variability, the planet's energy imbalance, and global climate forcings.

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 (Fig. **7b**), for which the ocean's thermal inertia minimizes the effect of weather noise. The cool anomaly of 2008 coincides with an ENSO minimum and does not imply a change of decadal temperature trend.

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 is confirmed by increasing ocean heat

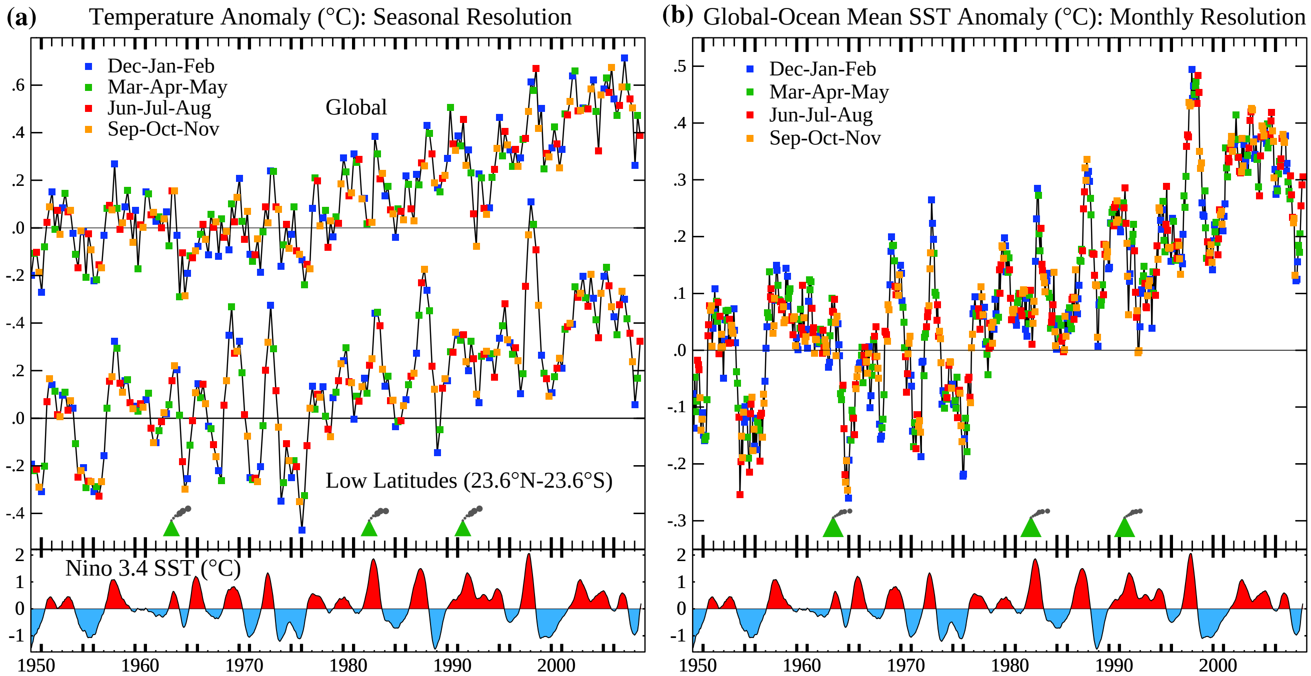

Fig. (7). (**a**) Seasonal-mean global and low-latitude surface temperature anomalies relative to 1951-1980, an update of [92], (**b**) global-ocean-mean sea surface temperature anomaly at monthly resolution. The Nino 3.4 Index, the temperature anomaly (12-month running mean) in a small part of the tropical Pacific Ocean [93], is a measure of ENSO, a basin-wide sloshing of the tropical Pacific Ocean [94]. Green triangles show major volcanic eruptions.

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 can 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 CO_2 within seven years (Supplementary Material section 18). Human-made aerosols cause a greater negative 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 human-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 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 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 CO_2 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 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. Although a case already could be made that the eventual target probably needs 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-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.

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