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ANALYSIS

Integrating economic analysis and the science of climate instability

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Abstract

Scientific understanding of climate change and climate instability has undergone a revolution in the past decade with the discovery of numerous past climate transitions so rapid, and so unlike the expectation of smooth climate changes, that they would have previously been unbelievable to the scientific community. Models commonly used by economists to assess the wisdom of adapting to human-induced climate change, rather than averting it, lack the ability to incorporate this new scientific knowledge. Here, we identify and explain the nature of recent scientific advances, and describe the key ways in which failure to reflect new knowledge in economic analysis skews the results of that analysis. This includes the understanding that economic optimization models reliant on convexity are inherently unable to determine an "optimal" policy solution. It is incumbent on economists to understand and to incorporate the new science in their models, and on climatologists and other scientists to understand the basis of economic models so that they can assist in this essential effort.

Keywords: Climate instability; Abrupt climate change; Methane hydrates; Clathrates; Dynamic Integrated model of Climate and the Economy (DICE); Economic optimization

1. Introduction

Early analysis of climate change by economists evaluated the transition between two climate equilibriums, or at most, two climate paths that smoothly changed over time from today's climate to one characterized by a doubling of atmospheric concentrations of CO_2 and other warming gases since the pre-Industrial Revolution (Mendelsohn et al., 1994; Manne and Richels, 1991; Gaskins and Weyant, 1993, and references therein). Cline (1992) was the first to extend the analysis beyond a doubling, with CO_2 emissions that were derived from simple models of economic growth (Manne and Richels, 1990; Nordhaus and Yohe, 1983; Reilly et al., 1987). Cline input the CO_2 emissions into a simple climate submodel that is used in the

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natural science literature for long-term analysis of the relationship between warming gases and temperature.[2] Cline then presented a sensitivity analysis of the benefits and costs of avoiding climate change with respect to the discount rate that converts future damages and costs into present values. Nordhaus (1992, 1994) developed an economic growth model (called "DICE") that endogenously calculated the interest rate, coupled with a sophisticated climate submodel, where the social discount rate used for benefit cost analysis of policy alternatives was based upon the endogenous rate of return on capital. Nordhaus' (1994) climate submodel is based upon a model published in the early 1980s that has the feature of climate equilibrium, in which an increase in CO_2 will eventually return to initial conditions (Schneider and Thompson, 1981).[3] While Nordhaus (1994, p. 26, note 4) acknowledged that his equilibrium climate submodel is not applicable to a greater than doubling of CO_2 equivalent gases, his model continues to be the basis for economic analysis beyond a doubling and is extended to analyze abrupt climate change. The science of climate change has advanced considerably in the last quarter of a century as these economic models were developed; yet the new understanding has not been incorporated in economic analysis of climate change.

Paleoclimatic and paleoceanographic studies completed during the past decade demonstrate that Earth has experienced dramatic and abrupt environmental change, at scales and rates not experienced during recorded history (Dansgaard et al., 1993; Mayewski et al., 1997; Alley et al., 2003). Scientific understanding of the nature of past climate change has been continuously advanced as climatic transitions are investigated at ever higher resolution and by application of new analytical techniques. These advances create a challenge for economic models in that specific climatic transitions that were only recently considered to be abrupt, but simple, steps of cooling or warming, have been discovered to actually consist of intervals of rapidly flickering climatic oscillation. Numerous rapid warming and cooling steps during the last 60,000 years have been 1/3 to 1/2 as large as the entire difference between the coldest glacial and warmest Holocene intervals, yet took only decades to years to occur (Alley et al., 1993; Alley and Clark, 1999; Severinghaus and Brook, 1999). Past decadal-scale increases in local temperature were more than 10 °C at high latitudes, as great as 7 °C at mid-latitudes, and 1–2 °C in the tropics (Alley and Clark, 1999; Hendy and Kennett, 1999; Lea et al., 2003). Both conventional and new hypotheses attempting to explain the forcing and amplifying mechanisms of such past climate instability have implications for future climate instability related to anthropogenic releases of greenhouse gases.

One broadly accepted explanation of climate instability invokes switching of the North Atlantic deep ocean thermohaline circulation that keeps Europe warm and distributes heat from the tropics to higher latitudes. A shutdown or slowdown could cause a step into glacial cooling (Broecker, 1997), whereas resuscitation of circulation patterns similar to those of today could produce rapid warming steps (Ganopolski and Rahmstorf, 2001). A more recent explanation is the "Clathrate Gun Hypothesis" (Kennett et al., 2003). Briefly, they hypothesize that relatively minor changes in thermohaline circulation warmed intermediate-depth ocean waters, causing instability of methane hydrates at depths of 400 to 1000 m, triggering collapse of continental slopes and massive releases of methane (a powerful greenhouse gas in short time frames) that reached the atmosphere. This hypothesis implies that future global warming events can be greatly amplified by the release of vast quantities of methane stored in the sea floor and arctic permafrost.

Past climate change was initiated by relatively small changes in the amount and distribution of solar insolation caused by cyclical changes in Earth's position relative to the sun ("orbital forcing") or by other, still undetermined, forcing agents such as variations in solar output, interplanetary dust, volcanism, etc. (Hays et al., 1976; Imbrie et al., 1992). The magnitude and rate of the resultant climatic changes, however, do not relate linearly to changes in the forcing function, because Earth's climate functions by stepping between a number of semi-stable operational modes of the connected atmosphere/hydrosphere/cryosphere system (Broecker and Denton,

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[2] See Hall (2001, p. 127) for a derivation of Cline's climate submodel, showing that it is equivalent to the climate model used by McElwain et al. (1999) to analyze climate at the Triassic–Jurassic boundary.

[3] See Nordhaus (1994, p. 33).

1989; Denton et al., 1999; Alley et al., 2003). Climatic flickering or abrupt transitions occur when thresholds in the combined ocean/atmosphere/climate system are reached. The critical values of these thresholds in terms of atmospheric composition, temperature, or oceanic circulation are yet unknown, yet abrupt transitions are features of many climate models dealing with ocean and atmospheric circulation (Manabe and Stouffer, 2000; Ganopolski and Rahmstorf, 2001). Anthropogenic emissions of greenhouse gases could provoke similar instability by forcing large enough change to reach such a climatic threshold, followed by cascading feedbacks that cause further instability and climatic flickering.

Economic analysis of climate policy is predicated upon the modeling assumption of a stable climate equilibrium at the present climate of the Holocene Epoch, an assumption that results in policy analyses concluding in favor of adaptation to climate change and against averting climate change. Environmental economists have not analyzed “climate instability”, although they have considered the possibility of a sudden climatic shift of known magnitude but unknown timing. Yet new scientific understanding is that climate change is frequently characterized by climate flickering and rapid, alternating changes of state that take place on the scale of decades or years (Alley et al., 2002). There is a broad consensus in the paleoclimate community on the timing, nature, rapidity, and geographic extent of these past abrupt climatic changes, in spite of substantial uncertainty and controversy about the triggers, thresholds, and processes involved. Thus far, economists have yet to examine the implications of true climate instability, including annual-scale climate flickering. Instead, economic models only admit a simple change in climate state in order to enable the calculation of an economic optimum. This limitation results in the policy conclusion to adapt to, rather than to avert, climate change.

2. The nature of past climate instability

Global climatic and environmental change has been documented over a wide range of time-scales by study of “proxy data” extracted from sediment and ice cores (Table 1). These data reflect quantitative and qualitative

Table 1
Sources of paleoclimatic evidence

This is a partial list of the most commonly used paleoclimatic and paleoceanographic proxy data, methods, and applications.

Ice cores from Greenland, Antarctica, and alpine regions: oxygen and hydrogen isotopes measure changes in temperature and seasonality of snowfall; trapped gases measure global concentrations of CH_4, CO_2, N_2O, and other gases; snow accumulation rates determined from annual layer thickness indicate humidity and storminess; amount and size of eolian dust and other particulates indicate aridity in source areas, windiness, and volcanic activity.

Marine sediment cores from continental margins and the deep ocean: microfossil assemblages (e.g., foraminifera, diatoms, and coccolithophorids), oxygen and carbon isotopes and trace elemental composition of calcite shells or alkenone ratios in organic matter measure the temperature, salinity, nutrient concentration of water masses and currents, global ice volume, and shifts in the global carbon reservoirs; pollen indicates vegetation on land; laminations and trace elements indicate paleo-oxygenation levels.

Lake sediments: pollen measures local geographical changes in climate and environment; microfossils (diatoms and ostracods), oxygen and carbon isotopes, and evaporite minerals reflect the balance of evaporation to precipitation, and the local carbon cycle.

Corals: oxygen isotopes and elemental ratios measure sea surface temperature and salinity, river discharge and rainfall on land and ENSO oscillations; age and position record sea level; cosmogenic isotopes record sunspot cycles; radioisotopes help calibrate the ^{14}C dating method.

Geomorphology and terrestrial geology: marine terraces record changes in sea level; glacial moraines indicate the extent of ice sheets; distribution and age of peat deposits record the extent of wetlands; seafloor pockmarks and submarine landslides record past methane venting events; eolian dust deposits (loess) records regional to global aridity; paleoshorelines of lakes record changes in evaporation/precipitation ratios.

Speleothems: oxygen and carbon isotopes and trace elemental composition of calcite in stalagmites and stalactites reflects changes in the hydrologic cycle and in temperature.

Tree rings: temperature and rainfall from ring width and density, and past variations in atmospheric radiocarbon abundance to calibrate the ^{14}C dating method.

Sources: Broecker (1997), Bradley (1999), IPCC (2001), Kennett et al. (2003), and the special issue of *Science* (27 April 2001, v. 292, pp. 658–659).

variations in atmospheric and ocean temperature and chemistry, vegetation, precipitation, etc. that have been calibrated by comparison with modern environments or derived from scientific principles (Broecker and Peng, 1982; Bradley, 1999). Fig. 1 shows climate variation on a number of scales, from longer Milankovich Cycles that reflect changes in the amount and location of solar insolation due to Earth's orbital variation, to shorter Bond Cycles and Dansgaard–Oeschger (stadial–interstadial) Cycles (Table 2).

Milankovich Cycles describe changes over 10's to 100's of thousands of years that result from variations in eccentricity (changes in the shape of Earth's elliptical orbit) every 413 ky and ~100 ky, obliquity or tilt (changes between 21° and 24° in inclination of Earth's rotational axis) every 41 ky, and longitude of the perihelion (precession of the equinoxes caused by wobble of Earth's spin axis) every ~23 ky. These variations controlled the major ice age cycles of the last few million years. The left column of Fig. 1, from a marine sediment core at Ocean Drilling Program Site 846, near Easter Island, shows the dominance of the climate record by the 100-ky cycle (warm interglacials marked: 1 = Holocene, 5(e) = Eemian, 7, 9, etc.) with the amplitude of each cycle increasing towards the present and the prior dominance of the 41-ky cycle before the last 800 ky. The center column of Fig. 1, derived from an ice core from Vostok, Antarctica, shows the characteristic rapid "terminations" of the glacial intervals, marked by steep temperature rises and overshooting in temperature at the beginnings of many interglacials, followed by a more gradual, but erratic decrease in temperature with a return to glacial conditions, producing an overall "saw-toothed" pattern of temperature variation.

The right column of Fig. 1 shows the glacial period of the last 90 ky, termination 1A at the beginning of the Bølling–Allerød (B–A) warming, followed by the cooling during the Younger Dryas, and termination 1B at the start of the warm Holocene. The last glacial period is marked by tremendous climatic instability, including ~1500-year Dansgaard–Oeschger Cycles (interstadials numbered IS 22 to 1), and the ~7500-year Bond Cycles. Bond Cycles also display a "sawtooth" pattern of stepped cooling that spans several Dansgaard–Oeschger Cycles, terminated by rapid warming immediately following the coldest episode. Although there are many globally recognized climatic events, past rapid climate change was neither everywhere synchronous, nor always similar in trend. For example, the Younger Dryas event was a dramatic ~1300-year-long cold period that interrupted the last deglaciation across much of the world, and was especially abrupt and distinct in the Northern Hemisphere. Fig. 2, however, shows an antiphase relationship between temperature in Antarctica and Greenland, where Antarctica experienced a cooling during the B–A that preceded the Younger Dryas. Fig. 2 also shows that the northern B–A warming has an overshoot in temperature followed by the sawtooth fall-off typical of many scales of global warming, and a cold climate flicker about 8.2 ky ago (8.2 ka) during the Holocene.

To understand the scale and rapidity of climate change, air temperature over Greenland changed 6° to 10 °C within decades or less during Dansgaard–Oeschger Cycles (Alley and Clark, 1999; Severinghaus and Brook, 1999; Kennett et al., 2003, p. 16). At lower latitudes, sea surface temperatures in the central North Atlantic (28° to 34°N) increased by 2° to 5 °C during interstadials (Sachs and Lehman, 1999; Kennett et al., 2003, p. 24), and up to 4° to 8 °C along the mid-latitude California coast in the northeastern Pacific Ocean (Hendy and Kennett, 1999). The tropics only experienced 1–2 °C increases at interstadials (Lea et al., 2003), but increased as much as 4 °C over the last full deglaciation. Such temperature changes are associated with shifts in precipitation and vegetation. Changes in vegetation (Kennett et al., 2003, pp. 58–61), increased windblown dust in ice cores (Broecker, 1997, p. 1584) and other evidence indicate that less rainfall created a dryer Earth during cold stadial and glacial periods.

Even the most rapid climatic transition appears to contain considerable climatic "flickering"—large, alternating jumps in temperature, precipitation, etc. that occurred on the scale of years. Fig. 3 shows the results of a detailed study of the fine-scale structure of a rapid climatic transition at the end of the Younger Dryas episode, approximately 11,600 years ago (Alley, 2000, after Taylor et al., 1997). The graph (progressing from right to left) shows that a substantial portion of the transition from the cold, dry, windy Younger Dryas into the warm, wet, early Holocene Epoch took place in about 25–35 years, yet contained substantial climatic flickers in the course of the transition. The

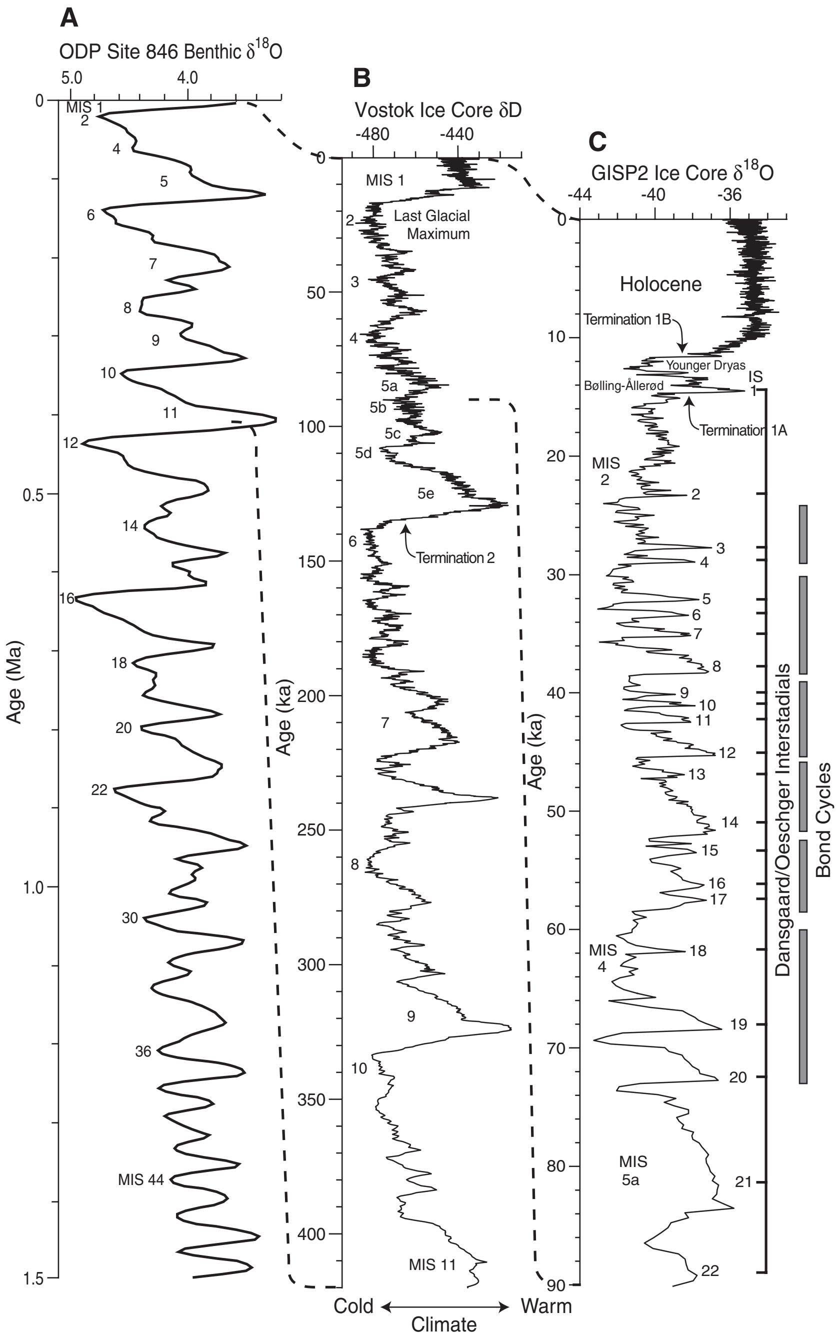

Fig. 1. Climatic oscillations and events of the last 1.5 million years shown with increasing detail with proximity to the present. Dashed lines connect each expanded interval. Warmer is always to the right. (A) Oxygen isotopic record from a marine sediment core showing the major glacial–interglacial oscillations at the scale of Milankovitch Cycles. Marine Isotopic Stages (MIS) are numbered, decreasing to MIS 1 for the Holocene Epoch (Shackleton et al., 1995; Mix et al., 1995). (B) Hydrogen isotopes from an Antarctic ice core showing the interval from 420 ka (MIS 11) to the present (Petit et al., 1999). (C) Oxygen isotopic record from a Greenland ice core, spanning the past major glacial–interglacial cycle, showing the millennial-scale variability of the past 90 ky (Grootes et al., 1993; Stuiver et al., 1995). Source: Kennett et al. (2003).

Table 2
Glossary of terms and acronyms

Albedo. Reflectivity of a surface; the ratio of the amount of electromagnetic energy reflected by a surface to the amount of energy incident upon it.

Archaea. An important and ancient group of prokaryotic microbes, distinguished from bacteria. This group includes methanogens, methanotrophs, extreme halophiles, extreme thermophiles, and the sulfate-reducing bacteria.

Benthic. Ocean or lake floor environments and associated bottom-dwelling life forms.

Biomarkers. Distinctive organic compounds, usually lipids, used to identify the presence of specific microbes or algal species.

Bond Cycles. Distinct ~7.4 ky oscillation with each cycle consisting of several stadial–interstadial episodes of decreasing duration or amplitude. Each cycle concludes with a severely cold Heinrich Event marked by an interval of North Atlantic ice rafting followed immediately by an abrupt warming into the next cycle.

Cenozoic. Geologic era extending from the end of the Cretaceous Period to the present, representing the last 65 million years.

Clathrate. Compound with a lattice of water molecules with gas molecules (e.g., CH_4, CO_2, H_2S) occupying cavities (or cages) within the lattice.

Cryosphere. Earth's ice, snow, and permafrost.

Dansgaard–Oeschger Cycles. Millenial-scale climate oscillations during the last glacial episode between warm interstadials and cold stadials (periodicity of ~1400 to 1500 years). Named after W. Dansgaard and H. Oeschger who first discovered them in Greenland ice core records.

Deglaciation. Interval of melting of the polar ice sheets following glacial maxima including glacial terminations and early interglacials. The last deglacial interval occurred from ~18 ka to 7 ka with particularly strong ice sheet decay between ~15 and 9 ka.

Foraminifera. Order of protists that construct a shell of calcite that can be preserved as a microfossil in sediment. A very important microfossil group for paleoceanography and biostratigraphy, especially for the Cenozoic Era.

Gas hydrates. See Clathrates.

Glacial terminations. Abrupt, punctuated episodes of major near-global warming and ice sheet melting that terminate glacial episodes. The last glacial episode was terminated by two such abrupt steps, Termination 1A at ~14.7 ka and Termination 1B at ~11.5 ka.

Glacial. One of many latest Pliocene through Quaternary climate intervals that occurred between interglacial episodes, marked by extensive ice sheet expansions and relatively cold climate over broad areas of the Earth.

Heinrich Events. Relatively brief (100 to 500 years) intervals marked by massive iceberg discharge from disintegrating ice sheets and associated meltwater episodes in high North Atlantic latitudes. These events are recorded as thick (several meters) sediment layers (Heinrich Layers) immediately prior to abrupt, major warming shifts of the last glacial episode.

Holocene. Most recent geologic epoch of the Quaternary Period representing the last ~11.5 ka of Earth history, corresponding to the most recent interglacial episode.

Insolation. The amount of solar radiation reaching a specific area of the Earth's surface.

Interglacial. One of many latest Pliocene through Quaternary climate episodes that occurred between glacial intervals, marked by relatively warm temperatures over broad areas of the Earth and of sufficient duration to have led to reduced ice extent much like that of the present day.

Intermediate waters. Cool water masses between the permanent thermocline and Deep Water (~400 to ~1500 m) produced in the Arctic and Antarctic regions.

Interstadials. "Short" warm climate episodes during the last ice age, each lasting hundreds to several thousand years, marked by glacial retreat and separated by cold, stadial episodes.

Intertropical Convergence Zone. Atmospheric division between northern and southern tropics forming a zone mostly north of the equator marked by convergence of surface (trade) winds in a region of lowest atmospheric pressure associated with warmest surface waters. Extreme upwelling of warm air in this zone leads to maxima in rainfall and cloudiness of convective origin.

ky. Thousand years duration.

ka. Thousand years ago.

Ma. Million years ago.

Methane hydrate. The ice-like state of CH_4 and H_2O, formed as CH_4 molecules captured within a cage of water molecules, produced under conditions of low temperatures, high pressure and sufficient gas concentrations. See clathrate.

Methanogenesis. The process of methane formation by microbes under anoxic conditions.

Methanotrophy. The process of methane oxidation by microbes.

Milankovitch Cycles. Astronomical cycles of climate change that resulted from fluctuations in the seasonal and geographic distribution of insolation caused by variations in the Earth's orbital elements: eccentricity, obliquity or tilt of the rotational axis, and longitude of the perihelion (precession). Named after Yugoslav mathematician M. Milankovitch (1870–1958).

Orbital-scale cycles. See Milankovitch Cycles.

Paleocene. First geologic epoch of the Cenozoic, following the Cretaceous Period and before the Eocene Epoch (~65 to 54 Ma).

Peat. Deposit of partially decomposed organic matter (largely plants) in a largely anoxic, water-saturated environment such as a bog.

(continued on next page)

Table 2 (*continued*)

Permafrost. Continental near-surface deposits that have been at temperatures below freezing for extended periods of time; perennially frozen ground.

Plankton. Pelagic organisms that float, drift or swim weakly. Generally very small to microscopic plants and animals.

Pleistocene. First geologic epoch of the Quaternary Period, prior to the Holocene (present), 1.8 million to 11,500 years ago.

Pockmarks. Crater-like depressions on the ocean floor commonly found on continental margins. Their diameter varies greatly from meters to 100 m with depths up to 10 m.

Productivity. The net rate of production of organic (biological) matter from inorganic sources of carbon in oceans or lakes, chiefly by photosynthetic plankton. Typically expressed in (mass of carbon assimilated)(area)$^{-1}$(time)$^{-1}$.

Quaternary. Interval of geologic time representing the last ~1.8 million years, consisting of the Pleistocene up to 11.5 ka and the Holocene since that time. The *late* Quaternary represents the last ~0.8 ky of this period and exhibits the largest climatic and glacial oscillations of the classic ice age.

Speleothems. Calcium carbonate mineral deposits formed by precipitation from water in caves (e.g., stalactities, stalagmites and flowstone).

Stadials. "Short" cold climate episodes during the last ice age, each lasting hundreds to several thousand years and marked by glacial advance.

Stratosphere. An outer layer of the atmosphere above the troposphere between ~10–50 km.

Thermohaline circulation. Motion of water in the deep ocean caused by density differences due to variations in temperatures and salinity and different from surface circulation which is wind driven.

Troposphere. The lower part of the atmosphere (up to 10–16 km) marked by rapid upward decrease in temperature, cloud formation and active convection.

Upwelling. The rising of cold subsurface waters toward the ocean surface. Since subsurface waters are often nutrient-rich, upwelling can lead to increased biological productivity at or near the ocean surface.

data show multiple 50–100% shifts in the rate of snow accumulation, some in less than 3 years, reflecting changes in humidity and storminess over Greenland. Variations in dust particle abundance, and electric conductivity chiefly indicate changes in windiness and the aridity of the dust's source area in Asia (Taylor et al., 1997).

Nearly all of the scientific understanding of rapid climate change and climatic instability is derived from investigation of the late Quaternary (past ~400 ky, and chiefly the last 100 ky). During this interval, Dansgaard–Oeschger-like changes, and even more rapid climatic flickers, shifted climate between glacial and interglacial extremes. There is relatively little knowledge of abrupt climate change during warm intervals, because the Holocene has been relatively stable compared to most of the Pleistocene (the early epoch of the Quaternary period). The Holocene is most similar to Isotope Stage 11 (Fig. 1, left and center columns), the longest-lasting (25–30 ky) interglacial of the past ~million years (McManus et al., 1999). This similarity is due to a nearly circular orbit of Earth, at present and during Stage 11, resulting in abnormally low amplitude variations in insolation due to Milankovitch Cycles (Loutre, 2003). In the absence of human impacts, the Holocene could last another 15–20 ky. This is not to say that warm, interglacial intervals are exempt from rapid change. The 8.2 ka rapid cooling event produced a dramatic change within the Holocene (Figs. 2 and 10) and resulted from freshening of Arctic and North Atlantic surface water—a very reasonable prediction for future global warming scenarios. There is no paleoclimatic information, however, from earlier in the Holocene or from prior interglacial episodes, on what is the likely response to continued increased greenhouse gas forcing (e.g., CO_2 and CH_4) during a warm interglacial. Although much larger in scale and occurring under very different conditions, the Paleocene–Eocene Thermal Maximum event, 55 million years ago, was a tremendous warming step that occurred on top of an already warm world that was likely related to a massive outgassing of CH_4 from methane hydrates below the seafloor. The release increased sea surface temperatures by 4° to 8 °C at high latitudes, and deep-water temperatures increased between 4° and 6 °C (Dickens, 2001; Norris and Rohl, 1999).

3. Conventional explanations of climate instability

Fig. 4 shows the concurrence over the last 420 ky of fluctuations in Antarctic temperature, and global atmospheric CO_2 and CH_4. A complete explanation of climate instability must account for this concurrence, as well as the fact that changes in solar insolation from

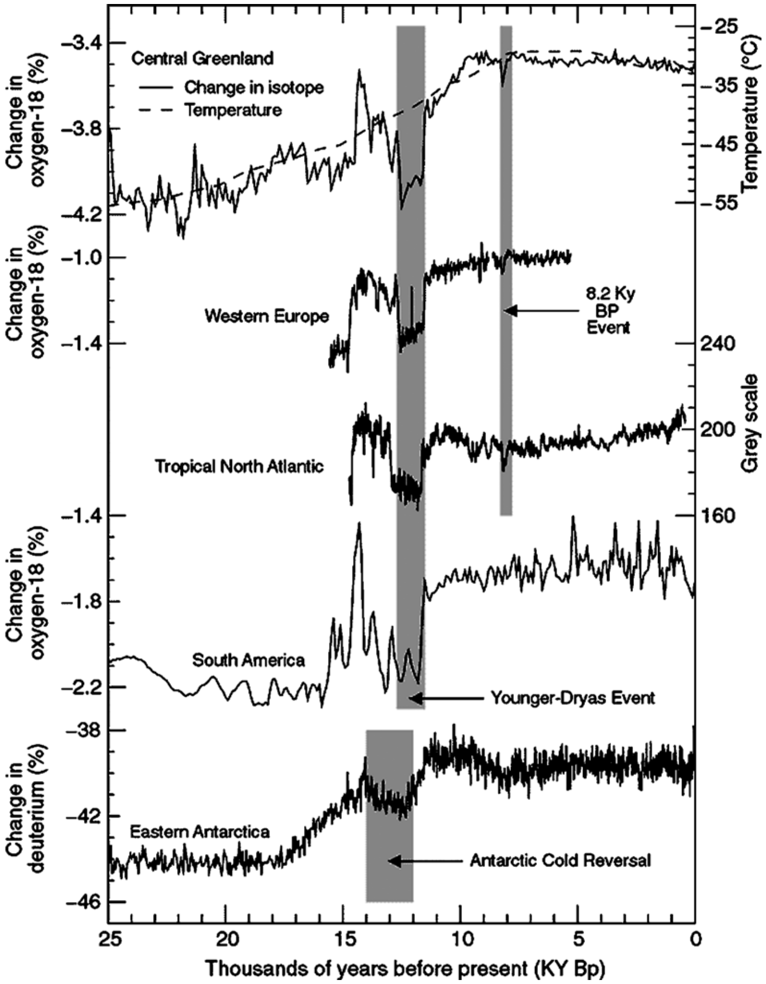

Fig. 2. Comparison of paleoclimatic records from four locations across the last deglaciation from 25 ka to the present. The Northern Hemisphere records (upper three) all show abrupt steps of warming or cooling at termination 1A (~14.7 ka), and the beginning and end of the Younger Dryas cold episode (~12.8 ka and ~11.6 ka). Temperature changes recorded by the southern hemisphere records are slightly offset, or even out of phase, with the northern hemisphere during the overall transition. Source: IPCC (2001).

Milankovich Cycles are not sufficient to explain temperature changes by themselves. The documented temperature changes require amplifying feedback mechanisms. Of these, changes in ocean thermohaline circulation play a key role in both conventional and new explanations.

Climate, atmospheric circulation, and ocean circulation are tightly linked in their role of transporting excess solar heat gained in the tropics to the poles where infrared radiation into space exceeds energy gained from incident sunlight. Three major atmospheric convection cells involved in this process determine the global latitudinal distribution of precipitation vs. evaporation, the locations of deserts and rain forests, and the mean strength and direction of the major bands of surface winds (Trade Winds, mid-latitude Westerlies, and Polar Easterlies). The evaporation–precipitation ratios and winds, in turn, control the distribution of salinity (and density) of seawater, and the direction and strength of surface and deep ocean currents. The meridianal movement of water, because of its high heat capacity and latent

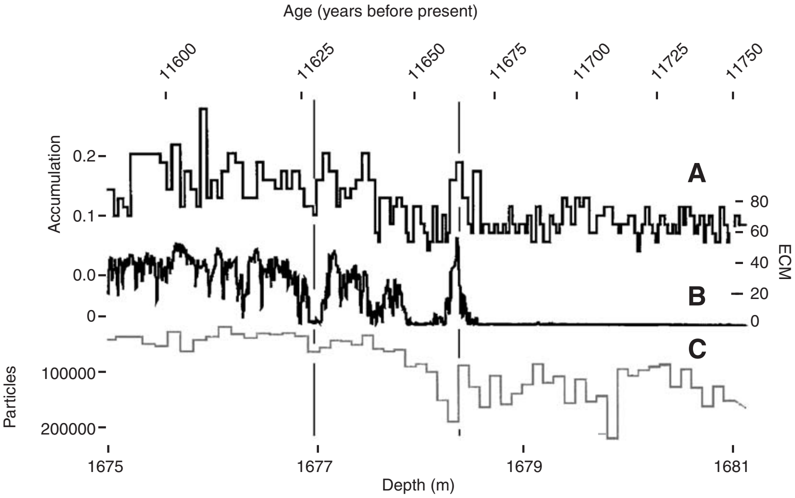

Fig. 3. The end of the Younger Dryas in the GISP2 ice core from central Greenland displayed at much higher resolution than Fig. 2. Data in upper curve displayed at approximately 1-year increments. The curves illustrate the "flickering" behavior through this step of the transition, wherein variables alternate between climatic extremes before stabilizing at the new equilibrium: (A) snow accumulation in m ice/year, reflecting humidity, precipitation and storminess over Greenland; (B) electrical conductivity (ECM), current in microamperes, reflecting acids in the ice core, possibly from volcanic dust; (C) the insoluble-particulate concentration in number per ml. The amount of dust in the ice reflects windiness and the aridity of the source area. The main step at the end of the Younger Dryas took place in ~25–35 years, between the two "flickers" indicated by the vertical lines at about 1676.9 and 1678.2 m depth. Modified from Alley (2000), after Taylor et al. (1997).

heats of evaporation and condensation, is particularly important in distributing tropical heat to the high latitudes.

In the North Atlantic, northward-flowing warm surface water becomes cooler and saltier as more water evaporates than accumulates through precipitation and run-off. The cold, dry atmospheric jet stream picks up latent and sensible heat as it flows across the northward-advecting water and carries it to northern Europe, moderating what would otherwise be a cold climate at that latitude. The combination of cooling and increased salinity increases the density of water reaching the Greenland–Iceland–Norwegian Sea, and it sinks into the deep ocean, creating a density-driven current (North Atlantic Deep Water) that travels southward, combines with cold water derived from the Antarctic shelves, and then proceeds to the Pacific and Indian Oceans, where it rises, creating a conveyor belt as the surface current returns to the Atlantic (see Fig. 5).

One explanation of climate instability is shutdown or variation in the rate of North Atlantic Deep Water (NADW) formation, (Broecker, 1997). Model simulations suggest that changes in temperature and salinity of the North Atlantic can abruptly shift the conveyor-belt thermohaline circulation between stable or semi-stable modes of operation (Ganopolski and Rahmstorf, 2001). There are many possible, contributing feedbacks in this system. As described above, temperature change between stadial–interstadial and glacial–interglacial periods is greater at higher latitudes and more moderate in the tropics. Greater warming at the poles would cause melting of ice and release of fresh water that reduces seawater salinity, reducing the density of surface water so that it no longer sinks, stopping or slowing the formation of deep water and shutting down deep ocean circulation. Heat would no longer be transferred to northern Europe and the high northern latitudes, where snow would have greater potential for extended preservation, thereby increasing albedo (light reflectivity). Sunlight would be more efficiently reflected instead of being absorbed by Earth's surface and converted to infrared radiation that warms Earth's surface and the troposphere. Extreme and rapid cooling events that had global to hemispheric-scale environmental impact were associated with the last major events of Arctic and North Atlantic seawater freshening 12,800 years ago during the last deglaciation (Younger Dryas) and 8200 years ago during the otherwise warm Holocene (the so-

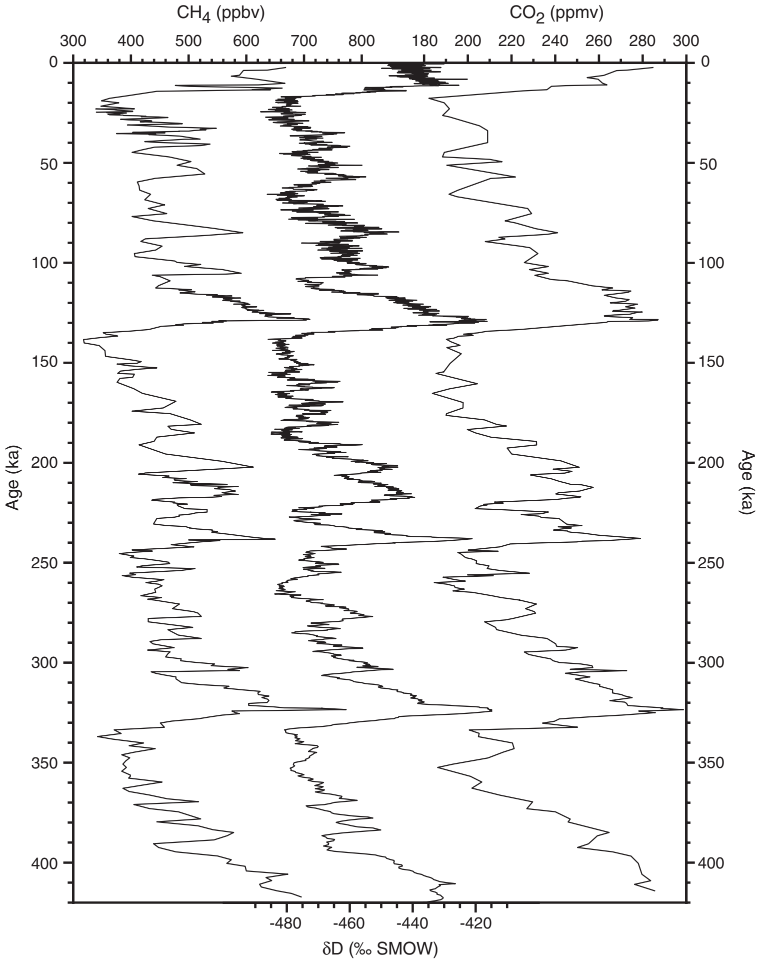

Fig. 4. Paleoclimatic record from the Vostok ice core, Antarctica of CH_4 (left), δD (middle) and CO_2 (right) for the last 423 ky. δD (hydrogen isotopic ratios) is largely a measure of the temperature of the atmosphere over Antarctica. CH_4 and CO_2 are greenhouse gases. Source: Petit et al. (1999).

called 8.2 ka event) (Alley et al., 2003). This explanation for the Younger Dryas cooling, in which climate change is driven or amplified by conditions in the North Atlantic, is consistent with observations of differences between climatic conditions recorded in Antarctic and Greenland ice cores and marine sedimentary records in the Atlantic Ocean (Fig. 2).

The ultimate trigger for warming events is not very well understood, but is conventionally explained as feedbacks driven by threshold-crossing variation in incident solar radiation modulated by Milankovitch Cycles. A slight increase in insolation results in more water vapor, the most significant heat-trapping gas, which amplifies the warming. Changes in other parts of the hydrosphere and the biosphere also influence climatic feedbacks, including atmospheric greenhouse gas concentrations. The appearance of wetlands results in terrestrial methane releases, another heat-trapping gas, and partially explains the concurrence between temperature and methane. The North Atlantic conveyer transports heat northward, and the ice sheets retreat, decreasing albedo and adding to the warming.

The major greenhouse gases that can be measured in ice cores (CO_2, N_2O, and CH_4) amplify any initiating warming signal from orbital forcing or solar variability: increasing during warm intervals and de-

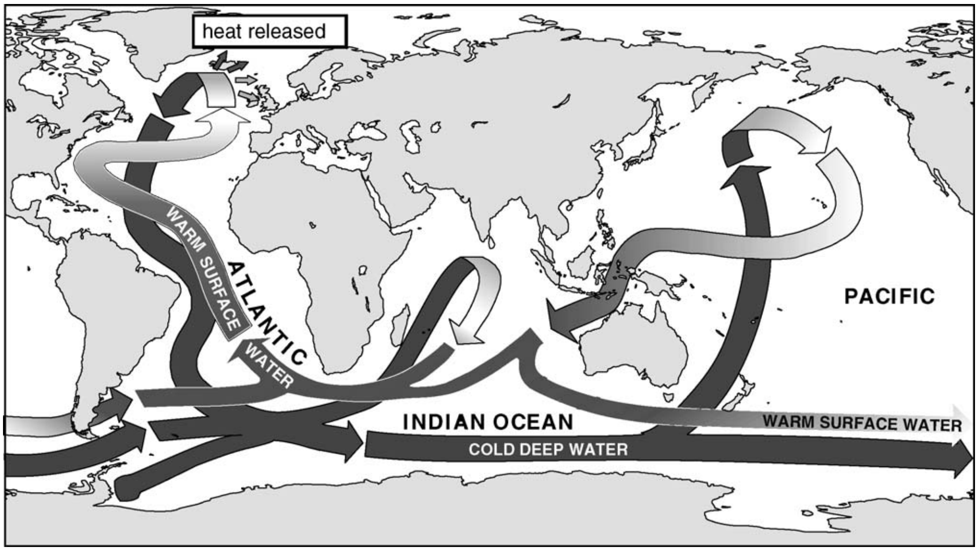

Fig. 5. The global "conveyor belt" showing a simplified version of the transport pathways of deep thermohaline and shallow water circulation. Transport of warm surface water to the North Atlantic to replace sinking North Atlantic Deep Water is a key element in the distribution of tropical heat to the high northern latitudes. Modified from Broecker (1991).

creasing during cold. At Milankovitch-scales, they covary with isotopic proxies of temperature. At a closer examination, atmospheric CO_2 appears to slightly lag glacial–interglacial temperature changes, and does not track Dansgaard–Oeschger Cycles during the last glacial period (IPCC, 2001, pp. 202–203). A combination of explanations for covariation of CO_2 with temperature is necessary (IPCC, 2001, p. 202, Box 3.4) and includes enhanced trapping of CO_2 in the deep ocean during glacials by changes in thermohaline circulation, increased wind-driven upwelling and phytoplankton productivity, and changes in ice cover. Somewhat similarly, the record of atmospheric N_2O has a smoother curve than temperature, following or slightly out of phase with the proxies for atmospheric temperature (Flückiger et al., 2004). This is consistent with gradually increased production of N_2O in wet forest soils and dry savannas during warmings (Kennett et al., 2003, p. 148). Water vapor is the most abundant greenhouse gas that amplifies an initial warming signal, but past levels cannot be directly measured from the gas trapped in ice cores. Sea surface temperature and a warmer atmosphere increase water vapor, especially in the tropics, as indicated by increasing monsoons during warming periods.

4. The Clathrate Gun Hypothesis

A major problem with the conventional conveyor-belt circulation hypothesis is that it does not explain spikes in atmospheric methane that are coincident with the onset of warmings, followed by saw-toothed decreases in methane that parallel falling temperature (Petit et al., 1999; Brook et al., 2000; Flückiger et al., 2004; Figs. 4, 6, and 7). For example, Severinghaus et al. (2003) find, "Atmospheric methane rose synchronously with temperature at this event,[4] within the uncertainty of our measurements." Methane is important because, at the short time scales of past rapid climate change – 10's to 100's of years, CH_4 has 23–62 times the greenhouse warming potential of CO_2 (IPCC, 2001). Conventionally, the fluctuation of methane has been solely associated with the extent of wetlands that expand after significant warming. Methanogenic bacteria (anaerobic Archaea) produce CH_4 by metabolizing and degrading organic matter, producing methane as a by-product. The methanogens cannot survive in oxygen, and are found in water-saturated soils or in sediment under water (fresh or

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[4] "This event" refers to rapid warming at the start of Dansgaard–Oescher event #8 (see Fig. 1).

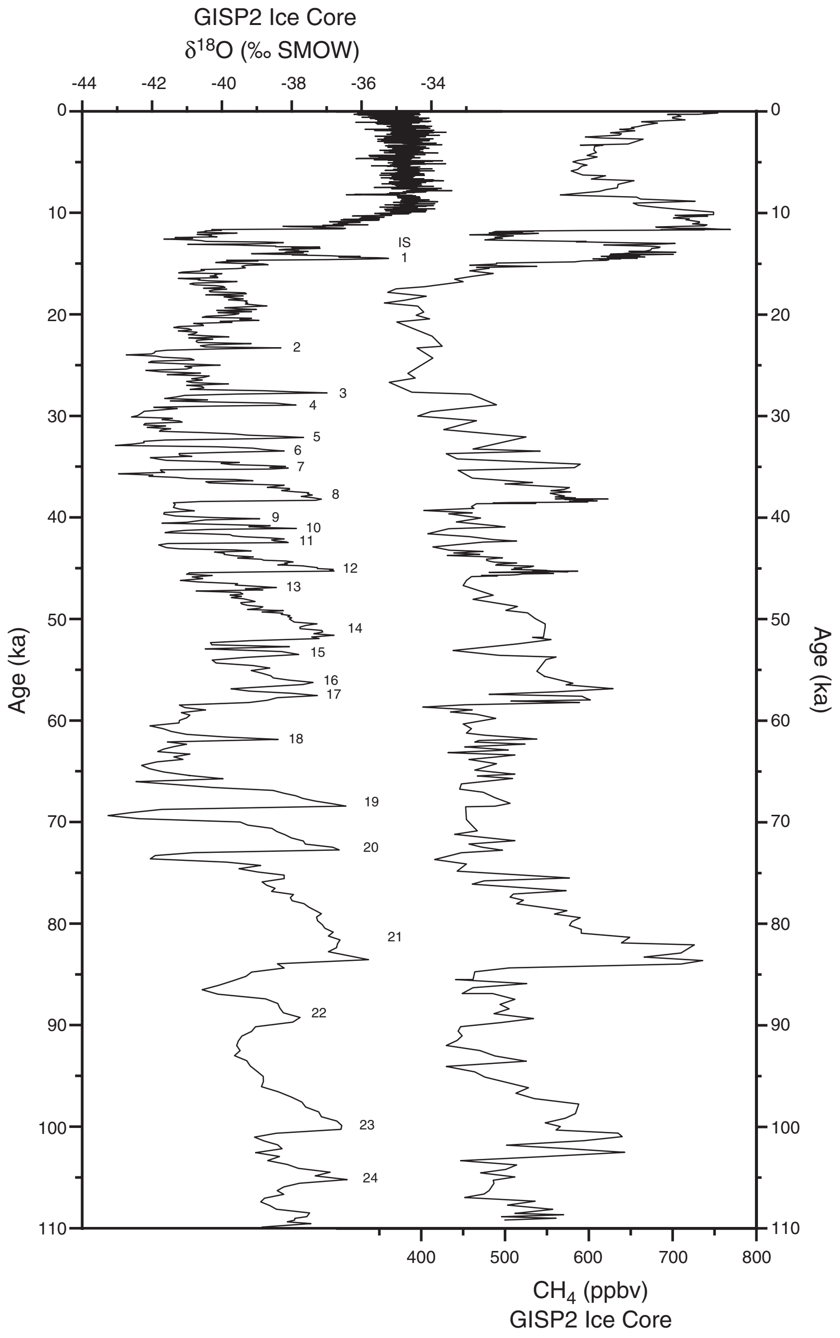

Fig. 6. Comparison of GISP2 δ^{18}O and CH_4 records for the last 110 ky from the Greenland ice core reveal a close correspondence of millennial-scale oscillations (Dansgaard–Oeschger interstadials) in air temperature over Greenland and atmospheric CH_4 concentration. Source: Kennett et al. (2003), based on data of Brook et al. (2000), Grootes et al. (1993), and Stuiver et al. (1995).

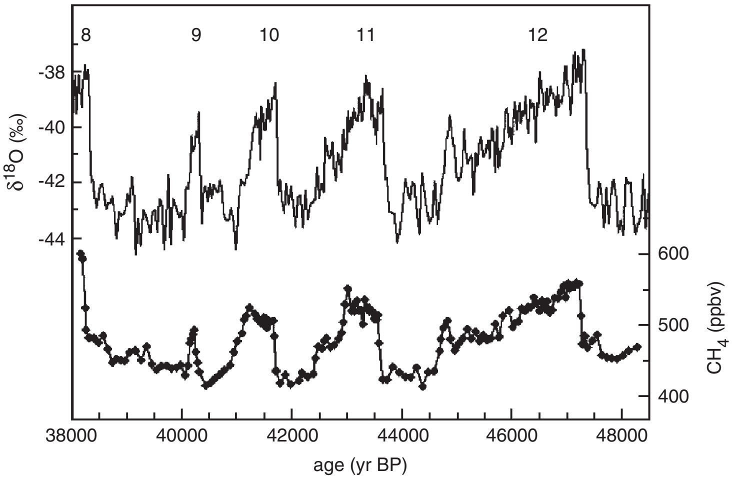

Fig. 7. NGRIP (North Greenland Ice Core Project) $\delta^{18}O$ (top), and CH_4 (bottom) for Dansgaard–Oeschger events 8 to 12, showing the exceeding close correspondence between air temperature and CH_4, in both timing and sawtooth character of the curves. Source: modified from Flückiger et al. (2004).

marine). The wetlands explanation is inconsistent with the CH_4 spikes that are coincident with the initiation of warm periods, and inconsistent with the saw-toothed drop to lower levels of methane that follows (Fig. 7), because wetlands only establish and gradually expand after sea level rises, flood plains become saturated, and organic matter can accumulate and degrade under anaerobic conditions. Kennett et al. (2003, pp. 43–87) review over 70 references of different types of data (see Table 1 for some examples) from locations around the globe that determine the timing and extent of wetlands relative to atmospheric concentrations. The data are inconsistent with the conventional explanation for wetlands as the source of methane because most methane-producing wetlands did not exist during the temperature and methane fluctuation of the Dansgaard–Oeschger Cycles and only became established significantly after the start of the Holocene.

An alternate source of methane – one capable of gradually storing and rapidly releasing vast amounts of the powerful greenhouse gas – is methane clathrate hydrates, also known as methane hydrates or gas clathrates. These are a form of methane–water ice that is composed of a lattice-like structure of water molecules with gas held in the cages of the lattice. Methane hydrates form in ocean sediment or permafrost at cold enough temperatures and/or high enough pressures to maintain a solid phase (Kennett et al., 2003; Fig. 8). If the temperature of water at the depth of the seafloor increases a few degrees, or if seafloor pressure is reduced by sea-level fall, existing methane hydrates will become unstable and dissociate into liquid water and gaseous methane (Fig. 8A). Fig. 8B shows that, even though the thickness of the zone of potential methane hydrate stability increases with depth in the ocean (deeper water is colder), there is a distinct base in the sediment below which the geothermal gradient has raised temperature beyond solid phase stability. Most methane hydrates are found in the shallower end of the stability zone, however, because there is a greater amount of methane-producing organic matter in sediments nearer to the continental margin. Methane gas from below the hydrate zone continuously travels upward into the uppermost sediments, forming gas clathrates in the stability zone. The methane hydrate-cemented sediment also acts as a permeability barrier, trapping free methane gas in the pore spaces of the sediment below, thus, in two ways, loading the “clathrate gun”. Methane hydrates are estimated to contain more carbon than any other reservoir of rapidly exchangeable carbon (i.e., oil, gas, coal), possibly twice as much as all other fossil fuels combined (Kvenvolden and Lorenson, 2001). This carbon is stored as the powerful greenhouse gas methane, the release of which is sensitive to changes in seafloor temperature or pressure (Fig. 8A).

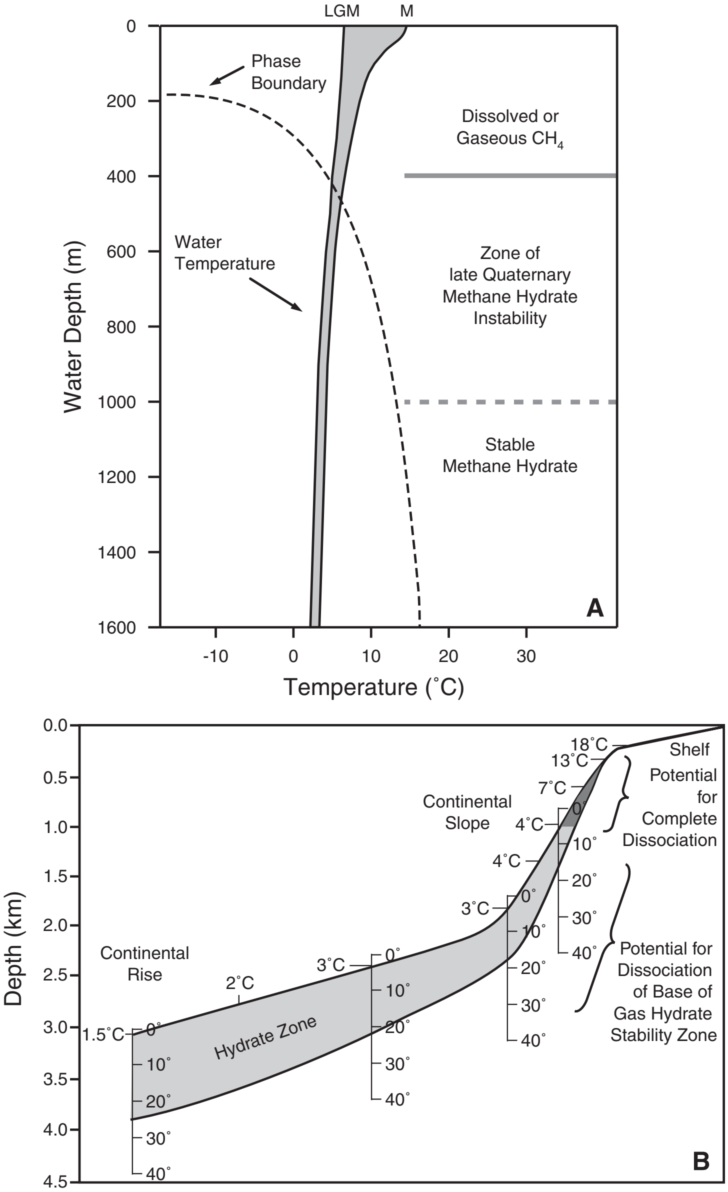

Fig. 8. (A) Effect of water depth (pressure) and temperature on phase relationships between methane hydrate and dissolved or gaseous CH_4. Solid methane hydrates are stable below and to the left of the curved, dashed phase-boundary line. Methane hydrates are unstable above and to the right of the line and would be dissociated into water and methane gas. Superimposed on this diagram (shaded area) is the range of southern California margin water temperature between modern (M) and Last Glacial Maximum (LGM) end members. Based on the known range of sea-level and intermediate water temperature variation during the late Quaternary, a zone of potential methane hydrate instability exists from ~400 to 1000 m where methane hydrates could be disrupted by reasonable fluctuations in temperature or pressure. Methane hydrates are increasingly stable and less susceptible to dissociation below 1000 m water depth. (B) Schematic diagram illustrating potential distribution and thickness of methane hydrate zone in continental margin sediments. Shaded zone represents potential areas of methane hydrate formation, where pressure and temperature conditions combine to produce methane hydrate stability, assuming a sufficient supply of CH_4. Thickness of this zone increases with depth (pressure) and lower water temperature. Geothermal gradient is assumed to be 27.3 °C/1000 m. Source: Kennett et al. (2003), after Kvenvolden and Grantz (1990) and Kvenvolden and McMenamin (1980).

Changes in sea level, and consequently pressure, during Dansgaard–Oeschger oscillations or other rapid climate transitions were too slow or of insufficient size to have caused methane hydrate instability and release on the time scales recorded in the ice-cores. However, recent discoveries of significant fluctuation in intermediate water temperature during stadial–interstadial oscillations provide a mechanism for destabilizing methane hydrates along the continental margins (Hendy and Kennett, 2003). Changes in thermohaline circulation or undercurrent strength at the depths of methane hydrate instability (~400 to 1000 m) occurred repeatedly in the Pacific Ocean during the last glacial period, with intermediate waters warming 1 to 2 °C, synchronous with, or slightly preceding, marine sediment and ice core indications of methane release (Hendy and Kennett, 2003; Kennett et al., 2003, pp. 113–124). Kennett et al. (2003, pp. 120–122) describe thermohaline processes that could trigger similar oscillations in intermediate water temperature in the Antarctic, Indian, and Atlantic oceans.

Higher seafloor temperature or lower sea level (lower pressure) can dissociate the clathrate structure into a gaseous phase, releasing methane gas into the water column. Aerobic methanotrophs generally consume the methane in seawater, except during large releases. Kennett et al. (2003, pp. 125–146) detail the seafloor morphologies of slumps, landslides, and explosive pockmarks associated with instability of continental slopes in the depth range of methane hydrate instability (~400 to 1000 m), as well as related deposits of disturbed and transported sediments. These show the dynamic nature of methane hydrate deposits and their temporal instability. Dissociation of methane hydrates

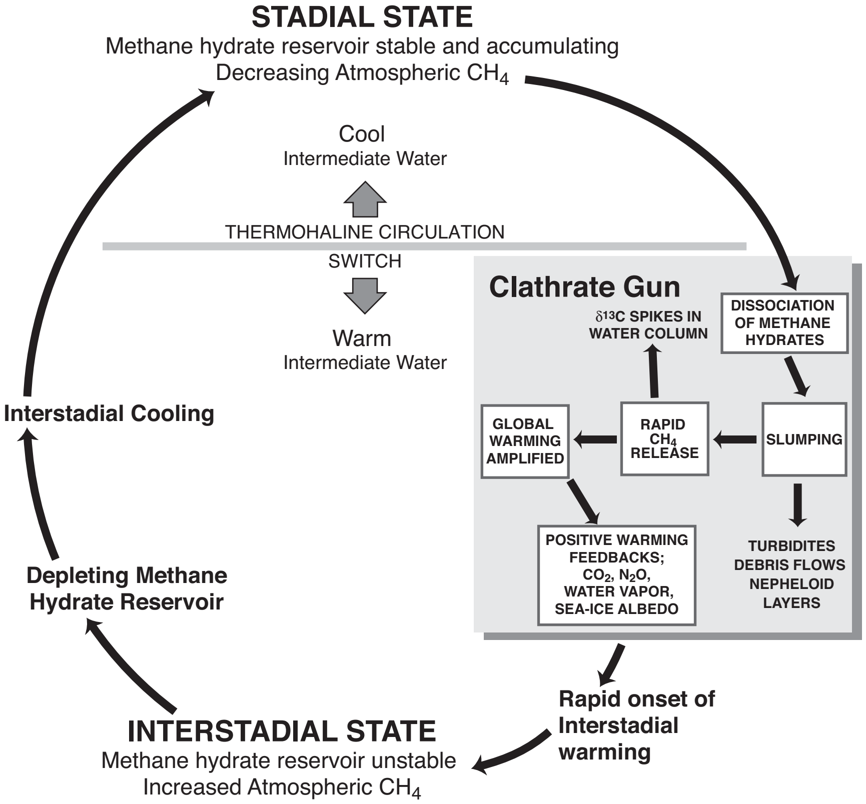

Fig. 9. Summary diagram illustrating major elements of the Clathrate Gun Hypothesis. In this hypothesis, climatic feedbacks associated with changing atmospheric CH_4 composition result from changes in the stability of the methane hydrate reservoir associated with changes in bottom-water temperature due to shifts in thermohaline circulation. Oscillations occur between stadial (glacial) and interstadial (interglacial) states. Source: Kennett et al. (2003).

increases the gaseous pore pressure in and beneath the gas hydrate zone, weakening it and contributing to submarine landslides and the eruption of gas that was trapped beneath. The locations and timing of known structural failures are largely consistent with a linkage between climate, atmospheric methane concentration, and the stability of marine methane hydrates (Maslin and Thomas, 2003; Maslin et al., 2004).

The Clathrate Gun Hypothesis (Fig. 9) suggests that a vast reservoir of potentially unstable greenhouse gas gradually accumulates (“loads”) then suddenly releases (“fires”) when intermediate waters cross a temperature threshold. Paleoceanographers trace the release of methane into the ocean by carbon isotopes and microbial biomarkers in marine sediments and by spikes in atmospheric methane at the initiation of rapid warming events. Although CO_2 is many times more abundant than methane in the atmosphere, on a 20-year time scale pertinent to the rate of past rapid climate changes, methane is 62 times more powerful as a greenhouse gas than CO_2 (IPCC, 2001). Furthermore, atmospheric methane is mostly oxidized within 20 years into CO_2 and water vapor, both of which continue to warm Earth. Episodic accumulation and discharge of vast quantities of methane, stored beneath the seafloor along the continental margins, provides another mechanism, in conjunction with shifts in the mode of thermohaline circulation, for rapid jumps and irregular flickers in Earth’s climatic state. The Clathrate Gun Hypothesis is compatible with co-occurring mechanisms that also release CO_2 stored in organic carbon and increase the flux of CO_2 from the deep ocean to the atmosphere due to deeper and more vigorous thermohaline circulation. Thus, as change in climate and ocean circulation pass key thresholds, abrupt shifts can occur in the integrated climate–atmosphere–ocean system that can be further amplified by separate, but related, feedbacks. Differences in the rate and phasing of these feedback mechanisms likely contribute to the erratic flickering during climatic transitions.

5. Equilibrium models for economic analysis

Climate instability has profound implications for the usefulness of models used by economists, as well as the analytical conclusions based on those models, such as that developed nations will benefit from modest amounts of warming. Equilibrium models have led to the general conclusion by economists in favor of policies to adapt to climate change rather than to avert it. This is surprising to us because this general policy conclusion in favor of adapting and against averting is promoted even in the context of acknowledging the real possibility of abrupt climate change (Nordhaus, 1999).

Ricardian models compare the economy in two equilibrium states, one with no climate change and the other with an equilibrium climate after a doubling of warming gases. Technological change, crop substitution, international trade, pest management, and other adaptations to a new climate equilibrium are integral to Ricardian models (Mendelsohn et al., 1994). These models inherently cannot assess ongoing climate change, much less climate instability.[5]

Most work by economists is based upon the impacts of a *doubling* of greenhouse gas concentrations on human and environmental systems. Since a doubling is likely to occur by the mid-21st century,[6] with much larger changes to follow, economic analyses should consider the consequences of larger and ongoing climatic alterations, especially in the context of new scientific understanding of past climate instability (i.e., stepping and flickering). In the following sections, we assess the weaknesses of dynamic economic optimization models for analysis of climate instability, considering climate submodels separately from the economic submodels.

6. Incompatibility between dynamic climate modeling by economists and recent advances in the science of climate change

More recent work by economists allows for dynamic changes in the economy, but assumes equilibrium in the climate portion that confines the analysis to a time frame up to a doubling of greenhouse gas concentrations, unless an exogenously given temperature replaces the climate portion of the model. In his influential work, Nordhaus (1994) presents the Dy-

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[5] For an alternative approach to assess the damage to agriculture from ongoing climate change, see Hall (1999, 2001).

[6] See Fig. 3.6 in Nordhaus and Boyer (2000, p. 62) and Fig. 2 in Hall (2001, p. 127).

namic Integrated model of Climate and the Economy (DICE). Nordhaus and Yang (1996) extend the economic portion of the models to a *R*egional dynamic *I*ntegrated model of *C*limate and the *E*conomy (RICE). Nordhaus and Boyer (2000) updated the DICE/RICE models. The original and updated models contain an economic model that calculates CO_2 emissions to the atmosphere and a climate model with two parts. The first climate submodel traces atmospheric concentrations of CO_2 as they are absorbed into the ocean.[7] The second climate submodel specifies atmospheric temperature as a function of atmospheric CO_2 concentration (and other GHGs), and traces atmospheric heat as it is absorbed into the ocean.

DICE and RICE contain several assumptions in the climate submodels, confining application to changes up to, but not extending beyond, a doubling of atmospheric concentration of CO_2 from the preindustrial level of about 280 ppmv. Nordhaus (1994) acknowledges two reasons why the submodels are not applicable for analyzing more than a doubling: (1) for greater than a doubling, carbon uptake by plants is limited, so that the atmospheric lifetime of carbon increases from the 120 years in DICE to between 380 and 700 years (Nordhaus, 1994, note 4, p. 26); and (2) for a 4-fold increase, Nordhaus acknowledges that models of Earth's ocean circulation switch to a new equilibrium, whereas in Nordhaus' model the ocean temperature eventually returns to that of today (Nordhaus, 1994, note 8, pp. 35–36).

There are other problems with DICE and RICE, making them inapplicable to modeling climate instability. The submodel of equilibrium CO_2 uptake by the ocean does not account for climatically induced changes in oceanic circulation, carbon reservoirs, or water properties. For example, the model ignores that with increased temperature comes decreased solubility, causing ocean degassing (Kennett et al., 2003, p. 148). The submodel also omits that CO_2 is variably released from or absorbed by the ocean due to changes in thermohaline circulation. For example, at the beginning of the B–A warm period, organic matter that sank to the bottom of the deep ocean during the last glacial maximum produced CO_2, which was released with upwelling of the deep ocean during large La Niña-like events between 13.8 and 15.7 ka (Palmer and Pearson, 2003). During past warm episodes (interglacials and interstadials), thermohaline circulation was deeper and more vigorous, increasing the CO_2 flux from the deep ocean into the atmosphere.

For the submodel of atmospheric temperature and heat transfer to the ocean, Nordhaus (1994, p. 40) states, "There is insufficient variation in the data (output from GCMs simulating warming from the pre-industrial period to 1990) to allow us to estimate more than two of the parameters, so we used physical data from the models to calibrate the two least important parameters." The "physical data" are based on a single number, the 500-year time that Nordhaus assumes for deep oceans (p. 37) to release one half of any additional heat; this is an equilibrium assumption inconsistent with the flickering nature of climate instability. Other physical parameters are values for the heat capacity of the top 133.5 m of ocean plus land and air, and the heat capacity of the deep ocean, considered to be only between 133.5 m and 1500 m in DICE/RICE, in essence treating depths below 1500 m as an unlimited heat sink. This is inconsistent with the recent historical record compiled and analyzed by Levitus et al. (2000), and inconsistent with theoretical results of coupled general circulation models of the atmosphere and oceans (Barnett et al., 2005). They show substantial changes in ocean temperature from the surface to 3000 m across the North Atlantic—all within the last 50 years. The "two least important parameters" are important for understanding even short-term climate change. In view of past fluctuations in intermediate-depth thermohaline circulation (Kennett et al., 2003), warming of intermediate waters is not uniform, and temperature increases along continental margins at intermediate depths could trigger the release of methane hydrates, further amplifying atmospheric temperature. In the case of thermohaline slowdown of the North Atlantic deep-water conveyer,

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[7] The original DICE and RICE models in essence specify two equations, one for the atmosphere and the second for the ocean, treating the ocean as an infinite carbon sink, a distinct limitation of the model. In the update, Nordhaus and Boyer change this into three equations, one for atmospheric CO_2 concentration, the second for surface ocean CO_2 concentration, and the third for deep ocean CO_2 concentration. The parameters, however, are specified so that less than one-half of 1% of CO_2 trapped in the deep ocean escapes per decade, an e-folding time of over 500 years (Nordhaus and Boyer, 2000, p. 60), so that the ocean is essentially a carbon sink within any relevant time frame of analysis, in contrast to the rates of change documented by the paleoceanographic and paleoclimatic evidence of Earth's history presented here.

thermohaline circulation in the Atlantic would be largely confined to intermediate waters with implications inconsistent with DICE/RICE.

The main problem with the DICE/RICE climate submodels is that the CO_2 concentration and the atmospheric/ocean temperature submodels are both equilibrium models with single long-run equilibria for CO_2 concentration and temperature, respectively. Compare a simple equilibrium shift with the erratic flickering at climate transitions shown in Fig. 3. To model climate instability, the climate submodels would have to admit oscillating changes in ocean circulation (both deep and intermediate water formation) with corresponding releases of methane to the atmosphere, CO_2 upwelling from the deep ocean to the atmosphere, and releases of heat from the ocean to the atmosphere. In addition, melting the permafrost will also release CH_4 into the atmosphere and change surface albedo. Moreover, various feedbacks amplify such effects. For example, as ocean surface temperature warms, warmer water does not sink as far, so the effective size of the ocean reservoir acting as a heat sink is reduced, yet DICE/RICE treats it as a constant. Climate submodels for economic analysis do not, but should, incorporate features that are consistent with the paleoclimate record and modern observations of the world oceans.

7. Climate instability and the inapplicability of economic submodels for economic optimization

The economic portion of DICE/RICE has four major deficiencies that make it inapplicable for modeling the economic consequences of climate instability: (1) the cost of emission abatement, (2) the treatment of the energy sector, (3) the damage function, and (4) a solution to the optimization problem does not exist. For the first and second deficiencies, the RICE/DICE models, and other "top-down" models, specify the cost of emission control as a reduction in the percentage of growth of gross domestic product. Top-down economic modeling assumes that markets already efficiently invest in energy efficiency and the technology to supply renewable and alternative energy sources. The result of this assumption is that policies that would speed the rate of technological change in these sectors are not considered. The economic literature is replete with critiques of the "top-down" approach, and we mention a few examples in a footnote,[8] but focus here on the remaining two deficiencies that are unique to economic optimization models.

In RICE/DICE, the damage function is equal for extreme warming or cooling, but does not capture the economic damage from climate flickering, nor the physical and economic destruction of capital caused by abrupt cooling or warming. Although the Holocene climate has been relatively benign and generally stable, the real possibility exists for anthropogenic release of CO_2 to trigger an otherwise unanticipated change in climate state (Alley et al., 2003).

Stocker and Schmittner (1997) model an increase in anthropogenic CO_2 and find that both the rate of emissions and the ultimate ambient concentration determine the shutdown of the Atlantic thermohaline deep ocean conveyer, an event that could result in climate change such as the ice age during the Younger Dryas. The ultimate threshold depends upon climate sensitivity,[9] and for one parameterization, the threshold varies between 650 and 700 ppmv (2.3 to 2.5 times the preindustrial ambient CO_2 concentration of 280 ppmv) depending on how slowly it is approached. This threshold will be easily achieved by economic activity in this century.

With abrupt climate shifts, such as an ice age caused by the termination or shift of deep water formation (Broecker, 1997; Stocker and Schmittner, 1997), we can expect destruction of location-specific functions of the ecosystem (Alley et al., 2003), and similarly we should expect destruction of the capital stock in location-specific regions of the world economy. During cold intervals (e.g., Younger Dryas and 8.2 ka events), permafrost extended over all of Canada, most of Northern Europe, and the Southern Island of New Zealand, for examples, and it was dry and windy in the tropics (i.e., Venezuela) (Fig. 10). During glacial periods, regional weather

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[8] Goodstein (2001) argues in favor of policy to encourage research and development of renewable energy sources, focusing on the path dependency of technological change. DeCanio (1997a,b) summarizes the top-down and bottom-up approaches. For analyses of the reasons why firms and consumers may not be efficiently investing in technology for energy conservation, see DeCanio (1997a,b, 1998, 1999), DeCanio and Watkins (1998a,b), Decanio et al. (2001), and the references therein.

[9] See Fig. 1 in Keller et al. (2000, p. 727) for a display of values as a function of climate sensitivity.

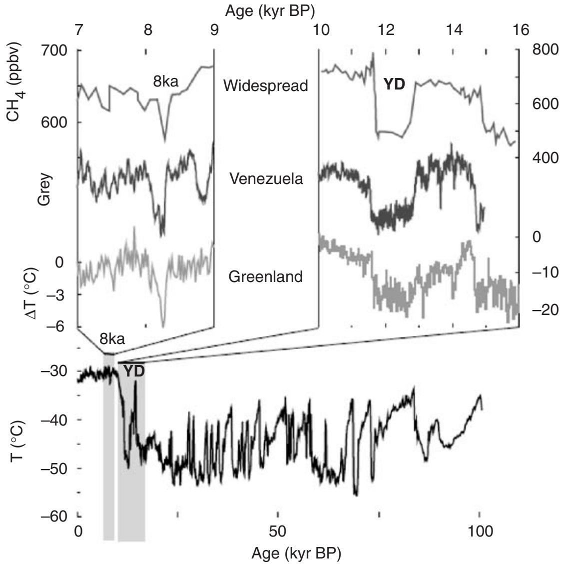

Fig. 10. Paleoclimatic data showing abrupt climate changes since Last Glacial Maximum. The lower panel is the history of temperature in central Greenland over the last 110,000 years (as in Figs. 1 and 6). Details of temperature for the Younger Dryas (YD) event and for the 8.2 ka cold events are shown as deviations from the temperature averaged over the intervals from 7000 to 8000 and 8400 to 9000 years ago. Methane concentrations reflect global production from all sources. Grey-scale shade of a sediment core from the Cariaco Basin, offshore Venezuela, reflects compositional variation in the sediment core; values are plotted here so that a downward shift corresponds to the effects of stronger winds over the basin or decreased rainfall on adjacent land. Source: Alley et al. (2003) after Alley et al. (1997), Brook et al. (1999), and Hughen et al. (1996).

patterns were organized much differently from today, for example, the jet stream brought moisture much further south, and the southwestern portion of the United States was much wetter than today, increasing water supply, but also increasing rates of erosion and sedimentation, which would consequently decrease the lifetime of dams and reservoirs. As climate warms, the frequency, magnitude and intensity of local weather events, such as hurricanes, monsoons, and droughts threaten water systems, flood zones, harbors, locations of orographic uplifting, coastal cities and low lying islands. Abrupt climate change, both warming and cooling, would result in physical and economic destruction of the capital stock of investments in hydropower and irrigation systems for agriculture and urban water use. The destruction of portions of the capital stock in the global economy would abruptly alter the rate of return on capital investment, and the rate of return on investment would be further altered in a discontinuous manner with each climate flicker and the expectation of additional climate flickering. The damage function should capture the discontinuities in changes of the capital stock, as well as associated discontinuities in the rate of return on capital.

In RICE/DICE, the damage function is symmetric for warming or cooling, relative to the absolute temperature today. A damage function that captures the essence of climate flickering should depend on both the rapidity of change and on the temperature relative to the recent past, for example as a geometric lag, rather than specifying zero damage with the temperature of today. The destruction of capital from climate flickering should change the expected return on capital investments. The cost of adapting to climate change should increase with flickering taken into account. The value of technological change to adapt to sudden decreases in temperature, precipitation and ambient CO_2, for example, should be lost with subsequent sudden increases in temperature, precipitation, and ambient CO_2.

There is a fundamental problem with economic optimization models like DICE for economic analysis of climate instability: a solution does not exist. Economic optimization requires convexity, and climate instability results in non-convex optimization functions. For example, DICE calculates the optimal capital stock over time and the optimal consumption stream. From either of these, there is an implied, internally consistent discount rate that should be equal, whether derived from consumption or from capital. A globally (in the mathematic sense) optimal path may not exist, so the discount rate must be imposed or calculated through some other means, for example by using the wrong counterfactual *such as a world without climate instability or no destruction of capital stock*—but then the model is no longer optimizing (see Keller et al., 2000).

Keller et al. (2004) use DICE to analyze the possibility of collapse of deep water formation by thermohaline circulation that would cause the climate to shift to an ice age. Their work is based upon a switch between climate extremes, calculated by Stocker and Schmittner (1997) to occur when RCO_2 equals 2.3 (650 ppmv) if CO_2 is increasing rapidly like today, an atmospheric concentration that we could reach before the end of this century.[10] If the damage from an ice age in percentage of GWP is lower than the cost of abatement, their model selects an ice age as an optimum,[11] results that are an artifact of the values of the parameters.[12] While their analysis is one of the first[13] to analyze an abrupt change in climate, it remains deficient because of a variety of limiting assumptions, such as a shift from the stable climate of the Holocene to a stable ice age, rather than to climate instability. Their assumptions include: (1) no destruction of capital; (2) replacing an estimate of damage with a simple assumption that a constant percentage of gross world product (GWP) would be lost; (3) a discount rate exogenously derived from a steady climate scenario that ignores the abrupt change; and (4) a temperature perturbation exogenously imposed on the model. These assumptions are inconsistent with the recent advances of climate science we have reviewed here that show that rapid, and irregular, climatic flickers are characteristic of climate transitions.

In the political debate today (2005) in the United States, there is a strong current of denial that Earth's warming is induced by the use of fossil fuels, as well as other anthropogenically caused releases of greenhouse gases. In one scenario analyzing the value of information, Keller et al. (2004) assume that policy makers learn in year 2085 exactly what the climate sensitivity parameter equals (either 2.4 °C, 3.6 °C, or 4.8 °C—see note 9). Since Keller et al. use the Nordhaus climate submodel that contains a smooth time path for global temperature, it is inconsistent to assume when policy makers will overcome denial (in year 2085) independently from how fast Earth warms (climate sensitivity). With their smooth growth in temperature, a more rapid warming rate should result in faster learning by policy makers. If the path of climate change is not smooth, however, it might be reasonable to assume that U.S. policy makers continue to deny the connection between anthropogenic release of greenhouse gases and climate change, a concept that Ha-Duong et al. (1997) call "socioeconomic inertia".

Over the last 100 years, ambient temperature increased in two distinct steps, fluctuating around a stable mean from 1860 to 1910, rising from 1910 to 1940, fluctuating around a stable mean from 1940 to 1980, and rising thereafter. A previous criticism of climate modeling was the predicted but unobserved continuing temperature increases, conditional on continuing increases in greenhouse gases from the 1800s to the present as observed in ice cores. While criticism and skepticism is central to the scientific method, this criticism has remained part of the denial today that climate change is anthropogenically induced. Levitus et al. (2000) compiled data on ocean heat content between the 1950s and 1990s and demonstrated that the ocean

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[10] RCO_2 is the ratio of atmospheric CO_2 concentration to the preindustrial level, usually set at 280 ppmv, but set in DICE equal to 290 ppmv. So in DICE, RCO_2 equals 2 with an atmospheric concentration of 580 ppmv (rather than 560, so DICE delays damage from emissions). Climate sensitivity is the increase in mean global temperature that occurs with a doubling, i.e., when $RCO_2=2$. The IPPC sets climate sensitivity to vary between 1.5 and 4.5 °C. Stocker and Schmittner (1997) consider climate sensitivity values between 3 and 4, while Keller et al. (2004) use 3.6 as the mean and $\pm 1°$ as a standard deviation for their sensitivity analysis.

[11] The value of the parameter b_1 in (6) of Keller et al. (2004), given in Nordhaus (1994, p. 21), limits damage from an ice age to 6.86% of GDP, while the cost of averting an ice age is allowed to be higher.

[12] Such a "result" is dependent on a large number of neoclassical economics-type assumptions, some less transparent than others. It is apparent, however, that if the maximum damage from climate change is simply selected to be small while the cost of averting is high, then a sensitivity analysis will always conclude in favor of adapting to climate catastrophe rather than averting it.

[13] For another recent effort to incorporate catastrophic climate change in an integrated assessment model, see Howarth (2000).

warmed during the pause in the rise of atmospheric temperature. Thus, the historic record of combined atmospheric and ocean temperature data are consistent with predictions by coupled general circulation models of air and ocean currents of surficial temperature changes caused by past anthropogenic greenhouse gas releases (Levitus et al., 2001; Barnett et al., 2001, 2005; Hansen et al., 2005). Hysteresis in atmospheric and ocean temperature is consistent with denial until some future year, such as 2085, of anthropogenically caused climate change (Hall, 2001). Learning depends on the magnitude and rate of climate change, but there may be significant lags in Earth's climate system that delays learning.

Our focus is on how climate instability can create non-convexities (with the result that an optimal economic solution may not exist) and how climate flickering can cause destruction of capital (with implications for any endogenous determination of interest rates). Other consequences of abrupt climate change undermine the traditional economic optimization approach: A "climate catastrophe" would create conditions so far outside contemporary human experience that there is no way to estimate empirically the utility loss associated with the catastrophe (DeCanio, 2003); There is disutility associated with the existence of the risk, even if the undesirable outcome never comes about (Nordhaus and Boyer, 2000).

8. Conclusion

The scientific understanding of the nature of climate change has gone through a revolution in the past decade with the discovery of numerous past climate transitions so rapid that they would have previously been unbelievable to reputable scientists. Reconstructed from a broad base of well-documented and consistent paleoclimatic records from around the world, large magnitude climatic transitions are now known to have been rapid (years to decades), large (3° to 10 °C over much of Earth's surface), and unpredictably irregular (climatic flickering oscillation, up to 1/2 as large as the entire transition, over periods as short as years). These transitions involved large stepwise shifts in surface and deep ocean circulation, atmospheric temperature, winds, aridity, and in the composition of powerful greenhouse gases that act as amplifying feedback agents. Abrupt jumps in the concentration of the greenhouse gases CH_4 and CO_2 provide a mechanism for globally synchronous warming events whereas shifts in the stable mode of conveyor-belt circulation directly and strongly influence the climate of the circum-Atlantic Northern Hemisphere. This new paradigm refutes assumptions that future climatic transitions are likely to be simple, gradual, or moderate, or consist of a single, abrupt switch to a new equilibrium state. Numerous positive and negative feedback processes that are beginning to be understood contribute to the irregular flickering that has characterized many past climatic transitions. In order to be accurate and pertinent to policy discussions, models of economic analysis of future global warming scenarios must include the best and most realistic understanding of the nature of global change.

Climate instability has important implications for the applicability of some approaches to economic analysis. For economic optimization models with and without policy intervention, prices are endogenous, the most important of which is the interest rate. The endogeneity of the interest rate requires an equilibrium model to equate over time the rate of time preference for consumption to the rate of return on capital. The interest rate is intrinsically important as a basis by which we make economically optimal tradeoffs between present and future consumption. If the equilibrium condition is not met, for the rate of time preference to equal the rate of return on capital, then policy conclusions based on economic efficiency arguments are inapplicable because the model is not endogenously selecting the interest rate. With irregular flickering between climate states, characteristic of past climatic transitions, we would expect the destruction of capital stock. If the flickering is forced by human activity, then policy or lack thereof results in the destruction of capital stock and a discontinuity of the rate of return on capital, violating the equilibrium condition. Moreover, with anthropogenically induced climate flickering that destroys capital, ex post of capital destruction, there is no guarantee that the rate of return on capital is positive.

We conclude that reliable economic analysis of climate change will require two major alterations to economic analysis. First, economists need to understand and incorporate into their climate submodels the recent advances by paleoclimatologists and paleoceanographers. Second, climate scientists need to understand how economists are modeling damage from climate change, so that they can educate economists about the deficiencies in economic models.

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