Overview

The physical concept of climate forcing

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Since the beginning of the debate on global climate change, scientists, economists, and policy makers alike have been using 'climate forcing' as a convenient measure for evaluating climate change. Researchers who run complex computer models conceived the theoretical concept of climate forcing in the late 1960s (Charney Report, 1979). This overview describes the development and basics of the physical framework, as radiative energy imbalance in the atmosphere, inflicted by a perturbation in the climate system. Such disturbances and forced changes can alter processes in the climate system, which enhance or dampen the initial effects and thus introduce positive or negative feedback loops. With increased understanding of the nature of the climate system, this basic concept has become more complex and hence more difficult to interpret. The identification of additional anthropogenic disturbances, the interdependence of individual forcings, and difficulties to account for spatial and temporal variabilities of disturbances are only few issues that complicate the overall picture. Although numerous scientific studies exist that evaluate climate forcings by allocating watts per square meter values to individual forcings (Intergovernmental Panel on Climate Change (IPCC) reports, 2010), the actual number of publications that interpret the physical meaning of the climate-forcing concept remains surprisingly small. Here, this overview focuses on explaining to an interdisciplinary audience the physical interpretation of the concept, including its limitations. It also examines new developments, such as polluter-based emission scenarios, energy budget approaches, and climate impacts other than temperature change. *WIREs Clim Change*

INTRODUCTION

In 1992,[1] the United Nations Framework Convention on Climate Change (UNFCCC) first addressed the need for policies and measures to stabilize atmospheric concentrations of greenhouse gases (GHGs) 'at a level that would prevent dangerous anthropogenic interference with the climate system'. The following 1997 Kyoto Protocol formulated targets of such policies or measures in terms of 'carbon dioxide equivalents'. Emissions of six species or groups of GHGs: carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), hydrofluorocarbons (HFCs), perfluorocarbons (PFCs), and sulfurhexafluoride (SF_6) were specified. It is worth noting that forcings and effects of man-made atmospheric particulates (aerosols) and land-use change were already introduced. However, only the six major GHGs are listed in the Kyoto Protocol. The so-called 'multigas abatement' strategy was conceived to implicitly address the 'comprehensiveness and cost effectiveness' guidelines of UNFCCC. Comprehensive was interpreted as including other gases besides CO_2, and cost effective was interpreted as giving options, of which gases or groups of gases can be reduced. However, to do so, emissions of gases with significantly different properties needed to be weighted and compared objectively. An independent currency of 'CO_2 equivalence' was envisioned where the weighting of climate impact can be physical, in the sense of 'comprehensive', or economical, in the sense of 'cost effective'.

In 1990, the then newly founded Intergovernmental Panel on Climate Change (IPCC) introduced the physical concept of climate forcing in its First Scientific Assessment Report (FAR).[2,3] A perturbation (like increasing GHGs) of an initial climate state that leads to a radiative imbalance in the atmosphere and that initiates a climate response became the physical

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framework of the 'climate-forcing' concept. Within this framework, the strengths of the feedbacks (the processes that dampen or enhance the forcing) in the system determine the relationship of 'radiative imbalance' and 'climate response'. Here in this text and referring to the original framework, 'climate forcing' is strictly defined as the 'radiative forcing' and is hence used as synonym for radiative imbalance, albeit other authors define climate forcing more broadly by including nonradiative effects in the concept. The goal of the overview is to clarify these subtle but important distinctions and shed light on the proper usage of the framework, which over the course of the years became significantly more complex.

After its introduction and from the first 1990 IPCC report onward, 'climate forcing' and its derivative 'global warming potential' (Box 1) became the basic physical metric, on which policies, measures, and economical assessments have been relying on. Other physical metrics, i.e., global temperature changes, or economic metrics that assess mitigation or damage costs were also introduced in the course of the IPCC process.[4] The purely physical climate-forcing framework, however, proved to be invaluable, especially in formulating policy targets.[5]

BOX 1

GLOBAL WARMING POTENTIAL

Global warming potential is the so-called emission metric, a tool that compares the climate impact of a unit of gas emitted to the atmosphere in a pulse to the impact of a reference gas, i.e., CO_2. The definition of 'climate impact' depends on the purpose. Climate impact can be described in an economic sense (cost versus benefit) or physically as in global warming, hence with regards to the latter the name 'global warming potential' (GWP). The GWP was developed as a metric in the Kyoto Protocol, with which the multigas emission strategy of limiting anthropogenic climate change can be internationally and politically implemented. The underlining assumption is that global warming is proportional to climate forcing and that the GHGs are independent of each other. The formulation used by the consecutive IPCC reports relates the radiative forcing of the emission of a kilogram of gas to the radiative forcing of a kilogram CO_2.The radiative efficiencies (radiative forcing per unit of mass increase) are thereby multiplied by decaying factors, which are the time-dependent atmospheric abundances of these pulse emissions. The integration of the forcings over the so-called time horizon of usually 20, 100, or 500 years is important for the political implementation. The advantage of the GWP metric is the ease, at which various industrial emissions can be globally compared. Long-lived GHGs and their GWPs are listed in the last IPCC report.[6] For short-lived gases such as methane and ozone, the GWP can be difficult to estimate. For aerosol forcings, the framework of GWP is not suitable because of the importance of their regional distribution, the short lifetime, and the impacts of anthropogenic aerosols on clouds and rainfall formation.

In this article, the section on *Climate Forcing and the Physics of Climate Change* revisits the physical basics of the climate system, while the section on *Climate Forcing and Earth's History of Climate* provides examples of the history of climate that exemplifies the role of forcings and feedbacks. The section on *Climate Forcing Concept* explains the history of the climate—forcing concept as it emerged from the developments of numerical climate modeling. The author bases the conceptual framework and nomenclature on the IPCC reports and key papers in the literature, particularly a series of papers by Hansen, published from 1984[7] to present.[8] In the section on *Climate Forcing Agents*, the general concept is evaluated by introducing various natural and anthropogenic forcing agents their spatial distributions, lifetimes, and climate responses that are relevant for the current climate change debate. A detailed description of individual forcing agents is beyond the scope of this article and can be best found in the latest IPCC report.[9] Commonly used, specific climate-forcing definitions, alternatives, and derivatives, such as efficacy of climate forcing, are discussed in the section *Climate Forcing Definitions, Derivates, and Alternative Metrics*. The gradual shift in the scientific debate, from global warming to more general climate change, challenges the framework of the climate-forcing concept. The latest developments in climate science emphasize cycles of energy, carbon, and water. The section *Energy Budget Approaches* scratches on some of these ideas. This chapter is followed by the conclusion section.

CLIMATE FORCING AND THE PHYSICS OF CLIMATE CHANGE

In general, climate is the long-term (at least 30 years) mean seasonal reoccurrence of weather pattern

mainly described by temperature and precipitation. Climate change is consequently a long-term change in the average weather. Regional climates differ due to proximity to the ocean, vegetation coverage, geology, altitude, abundance of sunshine, atmospheric composition, and so on. Therefore, climatologists study not solely weather pattern but the entire climate system that encompasses the atmosphere, hydrosphere (oceans and land water masses), biosphere (living organisms), cryosphere (sea, land ice, and snow), and lithosphere (rocks and geological formations).

The climate system can furthermore be described in a physical, thermodynamic sense as a weather-generating heat engine driven by the Sun's radiative energy input and release of excess heat to space in form of thermal radiation.[10] The system is stable if the net energy exchange at the top of the atmosphere is globally in balance. Note the wavelength band and intensity in which radiation is emitted back to space depends on the effective temperature of the planet. Thus, the energy balance requirement defines the effective temperature of the Earth (about 255 K), which is also the physical temperature of the mean level of emission in the atmosphere in about 6 km altitude.[10] When the Earth experiences a radiative imbalance initiated by a perturbation or due to internal fluctuations, the climate system is forced to change until a new global equilibrium is reached. The duration of the Earth's energy imbalance, and thereby the climate response time,[11] plays an important role.[12]

The planetary point of view of the Earth as a thermodynamic climate system neglects the fact that radiative fluxes are quite variable in space and time. Since the 1980s, scientists have been able to directly monitor fluxes of absorbed solar (incoming minus reflected) and outgoing thermal radiation at the top of the atmosphere. Examples of such measurements are shown in Figure 1. Solar radiation has most of its energy in the visible spectrum (top) and is reflected by clouds, atmosphere, and surface. The Earth's surface, clouds, and atmosphere emit thermal radiation back to space in the infrared spectrum (bottom) and thus cool the Earth.

Major land–ocean contrasts and regions such as Amazon or Sahara are recognizable in Figure 1. The seasonal variations of vegetation, ice, and snow cover are observable in the monthly images of net radiation (see Supporting Information). Apart from the spatial variability, individual months are rarely in global balance as well. Even globally averaged, the net radiative fluxes varied by about 0.5 W/m^2 over the past few years.[13] These observed variations are due to internal fluctuations within the climate system.[14] El Niño is such an example of a global weather pattern that shifts from one state to the other over the course of a few years[15] and, thereby, creates internal radiative fluctuations at the top of the atmosphere. A main challenge for accurate climate predictions consequently is distinguishing between internal fluctuations and climate forcings on a regional to global and decadal scale.

CLIMATE FORCING AND EARTH'S HISTORY OF CLIMATE

Throughout its history, the Earth experienced variations of its boundary conditions that produced globally positive and negative radiative imbalances at the top of the atmosphere (see green ellipses in Figure 2). Consequently, the climate system was forced into different states. Small, but persistent, perturbations (i.e., planetary movement) over extended time periods can accumulate over centuries and trigger significant climate responses as much as large instantaneous disturbances (i.e., comet impacts). However, the energy of forcings alone is rarely enough for major climate shifts. Positive feedback processes amplify the climate responses of these forcings. Arrows pointing in both directions within the gray shaded area of Figure 2 symbolize the pathways of these feedbacks that connect perturbations (ellipses) with climate processes (rectangles).

Perturbations can be radiative by nature, like changes in the Sun's output, or can be more indirect, such as disturbances in lithosphere or biosphere that alter the composition of the atmosphere and initiate a radiative imbalance, hence climate forcing indirectly[16](Figure 2). Forcings can also be regarded as internal or external to the climate system (inside or outside gray area in Figure 2). The distinction is slightly arbitrary and depends on the considered time scale.

For example in geological timescales, the lithosphere is a key component of the climate system. Therefore, geological disturbances involving plate tectonics (right green ellipse in Figure 2) such as seafloor spreading, uplifts of large plateaus, formation of mountain ranges, and amalgamation of supercontinents can also be considered internal perturbations. These processes may expose more areas to chemical weathering and draw down carbon dioxide, which would lead to a negative radiative imbalance of the planet.[17] Methane can, on the other hand, cause a strong positive climate forcing when released from methane hydrates buried in sediments of deep lakes and marine continental slopes by rising continents[18] (see 'plate tectonics' pathways in Figure 2). Note that it is speculated that methane release may have forced

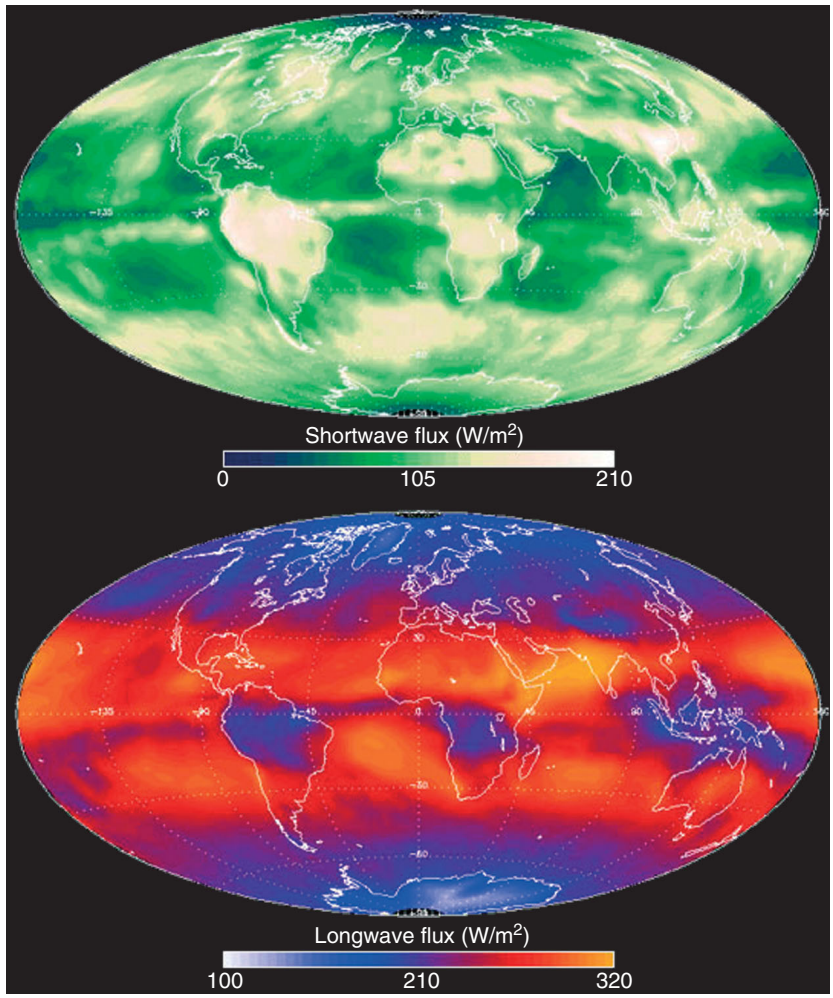

FIGURE 1 | Earth's shortwave and longwave radiation fluxes as measured from space. The two images show radiative fluxes in watts per square meters as measured from the NASA Clouds and Earth's Radiant Energy System CERES Instrument in March 2000. The shortwave flux measured by CERES is the portion of the radiative energy received from the Sun that is reflected back to space by the Earth's surface, clouds, and atmosphere. The incoming solar energy at the top of the atmosphere is measured by satellite instruments (not shown) and depends only on the solar output, the position of the Sun to the Earth, and the time and date. The absorbed solar energy of the Earth system is calculated as incoming minus reflected solar flux. CERES also measures outgoing longwave fluxes or thermal radiative energy emitted from the surface, clouds, and atmosphere as shown in the bottom image. The spatial patterns of these fluxes differ significantly albeit their global, long-term means have to balance in a stable climate. Outgoing thermal radiation is hemispherically symmetric, whereas continents strongly modify reflected solar radiation. (Reprinted with permission from NASA CERES Instrument Team (http://visibleearth.nasa.gov).)

the remarkable warming during the paleocene–eocene thermal maximum about 55 million years ago that resulted in a completely ice-free Earth.[16,19] In human timescales and in the current global climate change debate, however these forcings are considered external and thus placed outside the gray area in Figure 2.

The purest example of an external, radiative perturbation in all timescales is the variation in the Sun's activity. In geological context, for instance, during the Earth's Neo-Proterozoic about 500–700 million years ago, the Sun's luminosity was estimated to be about 6% lower than today. There is evidence in geologic data that in the Neo-Proterozoic the Earth experienced periods where the globe was completely ice covered and in a stable 'Snowball Earth' state.[20,21] The decline in solar output alone, however, cannot push the Earth in a snowball state. Water vapor and ice albedo feedbacks may have contributed significantly to this immense global cooling. Water vapor represents a change in atmospheric composition and ice growth is a nonradiative effect. Both processes feed back to radiative forcings as indicated in the pathways of Figure 2.

Other purely external perturbations are gradual variations of the movement of the planet around the Sun. The resulting solar changes are thought to force the climate system from glacial to interglacial climate states over periods of 100,000 and 22,000 years. Kepler's laws describe these movements of the Earth around the Sun with orbital parameters of eccentricity and obliquity. Eccentricity expresses the Earth's more or less elliptical orbit around the Sun, which ranges from nearly circular to slightly elliptical. Obliquity depicts the tilt of the Earth's axis perpendicular to the orbit of the Earth around the Sun with an estimated range from 22.1° to 24.5°. Note that obliquity is the reason the Earth experiences seasons and, therefore, higher obliquity angles enhance the seasonal contrast. Strong positive feedback mechanisms of the flora and fauna involving the carbon cycle and oceans and ice sheet dynamics must have been in place to account for these observed climate responses.[22]

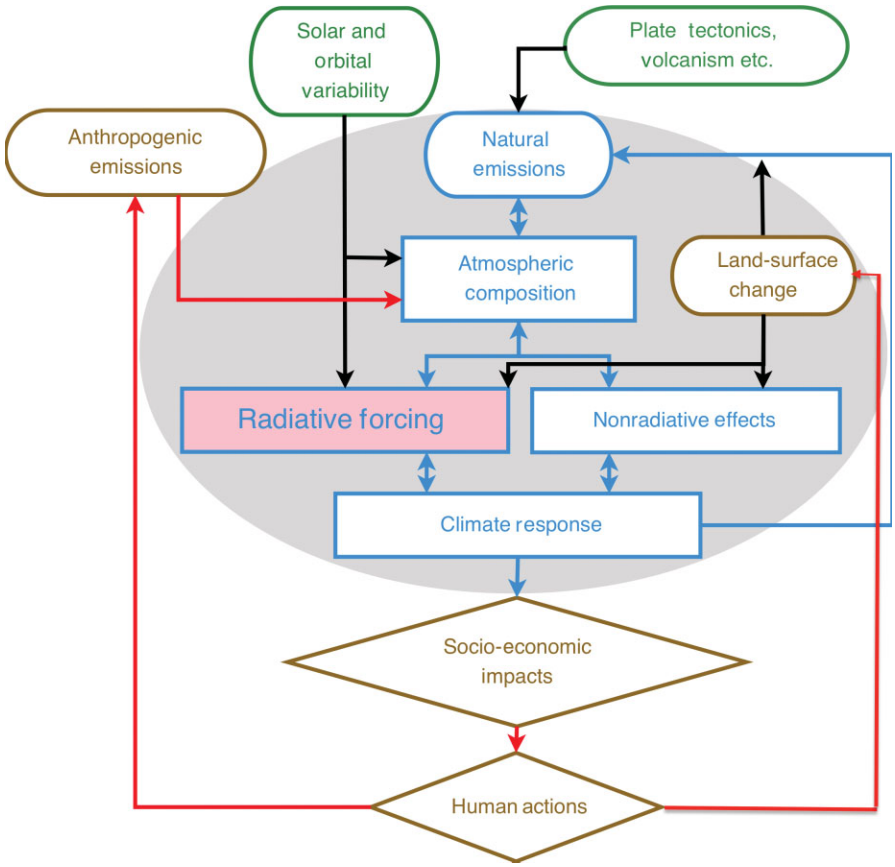

FIGURE 2 | The framework of climate forcing in the earth's climate system. This figure conceptualizes the impacts of internal and external perturbations on the climate system. Ellipses inside and outside the gray shaded area represent perturbations that modify (linked with black arrows) atmospheric composition, forcings, and effects. Feedback pathways, which couple physical and chemical processes (rectangles) of the climate system, are shown in blue arrows. Red arrows illustrate the pathways of human impacts symbolized by brown diamonds. In this diagram, the coupling of perturbations and climate response can either be through radiative forcing or nonradiative effects. Note in this illustration climate forcing is synonymous for radiative forcing, whereas other authors combine nonradiative and radiative effects for the climate-forcing concept. See text for further explanation of the diagram.

The variations in GHG concentrations are hence feedbacks of the orbital forcings as illustrated in the climate response–land surface change–natural emissions–feedback loop in Figure 2.

The current epoch of the Holocene began about 12,000 years ago with receding of the major continental ice sheets. Over this period, the Earth has been experiencing a fairly stable warm interglacial climate, which enabled the development of human societies. During the Holocene, orbital forcing modulated the climate by preferential heating of certain latitude bands and cooling of others. More intense plant growth sequesters more CO_2 and provides a strong negative feedback (see Ref 23 and citations therein and 'land surface' feedback loop in Figure 2). In subsequent centuries, atmospheric CO_2 increased again as shown in Greenland ice core records.[24] The centuries from about 800 to 1300 AD encompass the medieval warm period, which were warm and perhaps as warm as the present climate, which tree ring density data[25] indicate, and the Little Ice Age with extensive valley glaciers from around the seventeenth to the nineteenth century. One of the main open questions in the Holocene epoch is whether the extent of these climate shifts is global, hemispheric, or regional.

The illustration in Figure 2 furthermore includes the pathways of human interactions with the climate system and their coupling with the natural components that are of concern in the current climate change debate. Thus, the Nobel Prize winner Crutzen[26] coined the time period of the last 200 years the 'Anthropocene' epoch because of the beginning of early industrialization and in reference to humans who are now 'in control of climate'. Ruddiman[27] goes one step further and suggests calling the last 8000 years the Anthropocene epoch where humans were already living in all continents except Antarctica.

Ruddiman argues that humans have been modifying the landscape through deforestation and agricultural activity to the extent that climate can be impacted for thousands of years. These notions, although controversially debated, merit further readings.

CLIMATE-FORCING CONCEPT

Climate Forcing and Climate Sensitivity

Our knowledge of past climate variations and their driving factors stems from geological, geochemical, and paleo-climate observations. Some of these observations can be used as quantitative proxies for assessing the generalized response of the climate system to increasing GHGs. Radiative forcings of long-lived atmospheric gases trapped in ice, such as carbon dioxide and methane, are estimated from data of laboratory experiments of radiative absorptivity of these gases. For the same paleo-time period, past temperatures and their variations can be inferred from analyzing ratios of carbon isotopes of biological material or other isotopes of chemical elements that show a distinct temperature and/or precipitation signal. Furthermore, scientists estimate volcanic and solar radiative forcings from present day analogs and theoretical calculations based on Kepler's laws as described above. All this information is utilized to determine the general sensitivity λ of the climate system to perturbations such as changing GHG concentrations or solar forcing. One major issue that arises from paleo-climate studies is that radiative forcing RF and climate response may be linked by different feedbacks compared to present day climate. The strongest feedbacks, however, water vapor and ice albedo are dominant in all timescales (see Ref 8 and references therein).

When global temperature change ΔT is used as a surrogate for global climate response, the sensitivity λ can then generally be defined as follows:

$$\mathrm{RF} = \frac{1}{\lambda} \cdot \Delta T. \qquad (1)$$

The radiative forcing RF is the difference of the global radiative energy budget at the top of the atmosphere of two climate states, i.e., with two different orbital parameters or two CO_2 concentrations from paleo-observations. The linear relationship between forcing and global surface temperature difference of the two respective climate states determines the sensitivity parameter λ, which, in a general sense, describes the innate efficiency of the climate system as a heat engine. Specifically it describes the strength of feedback mechanisms that dampen or enhance the forcing RF. Radiative energy (forcing) is consumed by the system to perform work (as convection, evaporation of water, melting of ice, geochemical reactions, biophysical, biochemical, and ecosystem changes) and heat the system (to warm the atmosphere, land, and ocean). This relation of global radiative forcing and surface temperature change has shown to be surprisingly stable for paleo-climate studies. This means that the sensitivity parameter λ is more or less independent of the type of climate change drivers such as CO_2, volcanic, and solar forcings.[7]

Forcing Concept in Climate Models

Climate scientists also developed a theoretical framework of radiative forcing. The concept originates from early studies of climate responses to solar forcing and doubling of CO_2 forcing in the 1960s. Radiative transfer models with various fixed atmospheric and surface conditions were developed. The so-called radiative-convective models added nonradiative energy transfer and interactive adjustments in the atmospheric column.[28] In the 1970s, these studies were further extended to computer models with mathematical descriptions of atmospheric general circulation, and simplified representations of ocean heat transport hereby adding the third dimension. An historic account of these early developments is summarized in the famous 'Charney Report'[29] of the United States National Academy of Science from 1979 (see Ref 30 for historic background).

Figure 3 illustrates the solar and thermal radiative and nonradiative energy fluxes of dry convection and latent heating together with a typical temperature response profile in the atmospheric column. Thermodynamic and dynamic processes let the atmosphere react to an imposed radiative energy imbalance (climate forcing) by adjusting its entire vertical temperature profile to a new equilibrium (climate response) through the work of lifting and expanding of air, evaporation and condensation of water, and radiating excessive energy back to space (climate feedbacks). Radiative–convective models describe these weather-related adjustment processes within the troposphere—ocean/land system (troposphere is the atmospheric layer between the surface and approximately 10 km). In the stratosphere, the layer above the troposphere, temperature adjustments are small and mainly controlled by ozone photochemistry. As a consequence, the stratospheric temperature radiatively adjusts to perturbations within a few months, whereas adjustments of the troposphere—ocean/land system take years to decades. The degree to which adjustments are included in calculating the radiative imbalances determines the differences in the forcing

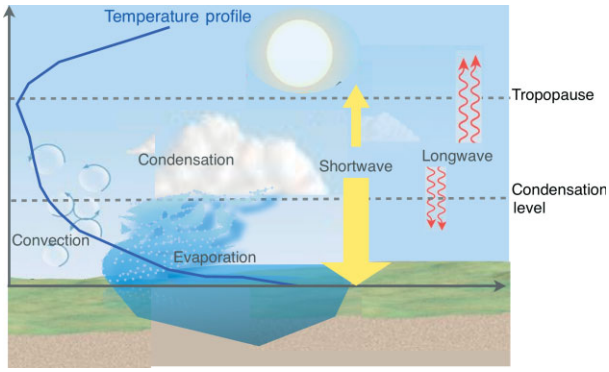

FIGURE 3 | Energy fluxes and atmospheric temperature profile. Illustrated (Figure 3) are the upward and downward solar (yellow arrows) and terrestrial (red arrows) radiative fluxes, the energy transport from the surface to the atmosphere via evaporation of water and convection of air, and the release of the evaporative energy in the atmosphere via cloud formation (condensation). The blue line represents a typical temperature profile from the surface to the tropopause and above as it evolves from these energetic flux adjustments. The positive radiative energy budget at the surface has to be balanced globally by evaporative and convective cooling with a small contribution from heat uptake of the oceans and land and melting of ice and snow. Convection describes lifting of warm air, which expands and cools. Cloud formation occurs as condensation of water from the cloud base. Outgoing terrestrial radiation from the surface warms the atmosphere due to clouds, water vapor, and natural and anthropogenic GHGs that absorb the radiative energy. The air and clouds emit thermal radiation back to surface and to space. Temperatures increase again in the stratosphere as indicated with the blue curve mainly due to absorption of UV-light by ozone.

definitions, as explained in the *Climate-forcing Definitions* sections.

Climate modelers often study temperature responses to global forcings by varying the boundary conditions of general circulation models (GCMs) and running the model until equilibrium is reached. The undisturbed case is, hence, called the control run. Then the model is rerun again under new, perturbed conditions (e.g., doubling of CO_2). The assumption is that the global mean surface temperature difference between the two final equilibrium states is the estimated temperature response of the climate to the imposed perturbation after the climate is allowed to adjust. The ratio of unadjusted forcing (calculated independently) and the temperature response determines the climate sensitivity parameter λ in Eq. (1).[31] Note that climate models are not always run to equilibrium but in transient mode.

Equilibrium Climate Sensitivity Versus Transient Climate Response

The classic GCM equilibrium studies investigate longwave forcing of doubling of CO_2 and shortwave forcing of 2% increase in solar irradiance.[7] The climate sensitivities of these two forcings turn out to be quite similar and hence are considered independent of the forcing agent. On the other hand, the differences in equilibrium climate sensitivity between individual models are quite large. Note the equilibrium climate sensitivity for doubling of CO_2 was estimated between 1.5 and 4.5°C in the Charney Report of 1979 and, three decades later, it is between 2.1 and 4.4°C in most models of the fourth IPCC Assessment Report.[6] GCMs that couple the atmosphere and the ocean can also be run in transient mode, which includes continuously changing perturbations. A classic example is the 1% CO_2 increases per year up to doubling of CO_2. In this case, the model does not reach equilibrium and the ocean takes up a considerable amount of radiative energy. Hence, after balancing the atmospheric energy fluxes at doubling of CO_2, the ocean remains a heating source and warms the atmosphere from underneath for some time. The response of a transient perturbation is often called transient climate response.

CLIMATE-FORCING AGENTS

The IPCC generally refers to the climate change drivers that impose perturbations to the climate system as 'climate-forcing agents". Climate change drivers can be ongoing external perturbations (e.g., solar forcing), perturbations with long or short lifetimes (e.g., gases or particulates in the atmosphere), and can be natural, man made, or a combination of both. Focus of the global warming debate is on the anthropogenic, long-lived agents that act on top of the continuously occurring natural perturbations, but short-lived anthropogenic drivers cannot be neglected either, when they show a long-term trend in emissions. Natural and anthropogenic forcing agents that are important for the time period of industrialization from about 1750s on are described in detail below. A full account of all known forcings from preindustrial to present day can be found in the latest IPCC report,[9] from which Figure 4 is taken.

Volcanic Forcing

Volcanic eruptions are climate change drivers that cause climate forcings on several timescales. On geological timescales, the amount of CO_2emitted might be small, but the fact that the degassing lasts over millennia can initiate major climate shifts. On the other hand, SO_2 gas and also ash particles injected in the atmosphere by explosive tropical volcanoes can reach levels up to 20 km where the

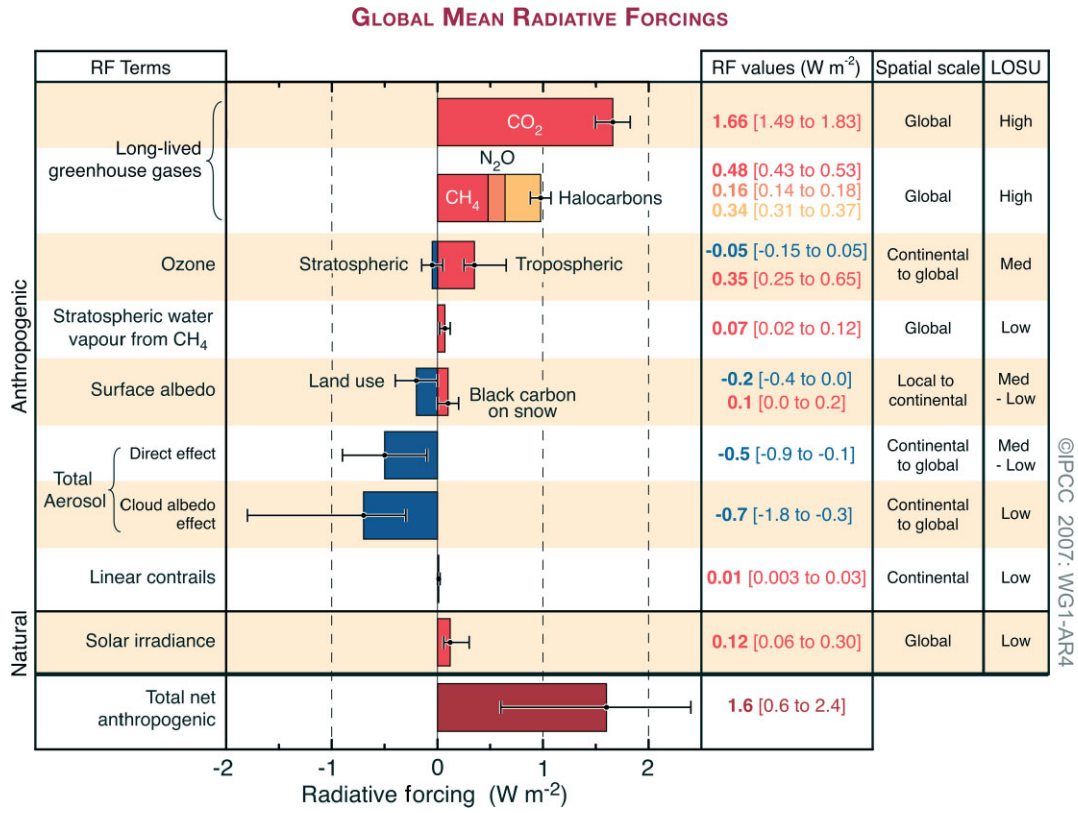

FIGURE 4 | Major radiative forcings of present day (2005) relative to the start of the industrial era (about 1750) after IPCC-AR4.[9] The terms or type of perturbations, the corresponding radiative forcing as histograms and listings including error bars, the spatial scale of the forcing agents, and the levels of confidence in the estimates given by the IPCC-AR4 are listed from left to right. Note that some perturbations, such as aerosols, exert nonradiative effects such as precipitation changes or changes of weather pattern, which may cause additional radiative forcings that are not accounted for in this figure. See text for further information.

gases can be transformed into tiny sulfuric acid drops (volcanic aerosol). Once in the stratosphere, the aerosol layer can remain present up to 3 years and can spread globally. The volcanic aerosol particles in the submicron size range reflect sunlight back to space and therefore affect the climate system negatively. Volcanic forcing may be the strongest short-term forcing. The peak global radiative imbalance at the top of the atmosphere was about -3 W/m^2 for the last major volcanic eruption of Mt. Pinatubo in 1991.[32,33] The climate effects of the Mt. Pinatubo eruption are well studied and include reduced global temperatures of about a half of a degree in 1992 and 1993 with increasing winter temperatures in Northern hemispheric polar regions and summer cooling in the tropics and subtropics.[34] Increasing CO_2 uptake of more productive vegetation due to more scattered sunlight was also observed.[35]

Solar and Orbital Forcings

Over the last millennium, the orbital forcing has been slightly positive and in the order of a few tenths of a watt per square meter.[36] For the present day climate, the relevant solar forcing is the 11-year solar cycle, which describes the variation of the intensity of Sun's output that changes with the occurrence of faculae (bright flares) and sunspots (dark spots) on the Sun's surface. These solar irradiance variations are only in the order of 1 W/m^2 or 0.1% of the total intensity of the sunlight at top of the atmosphere. However, faculae and sunspot variations are more 'visible' in the ultraviolet spectrum, reducing the intensity by about 1%. In the stratosphere, the dangerous ultraviolet part of the sunlight drives photochemical reactions of the ozone chemistry. Therefore, a relatively small solar forcing affects ozone concentrations (see below).

Forcings of GHGs

The IPCC traditionally divides GHGs into the categories: atmospheric carbon dioxide, methane, other Kyoto Protocol gases including N_2O, Montreal Protocol gases, and ozone.

Atmospheric CO_2

Biogeochemists distinguish between organic and inorganic, terrestrial and oceanic, and short- and long-term carbon cycles.[23] The short-term organic cycle consists of plants and marine microorganisms that remove CO_2 from the atmosphere through photosynthesis and store organic carbon in biomass where it is buried in soil and sediments. Aerobic and anaerobic decomposition releases the carbon again in form of methane and carbon dioxide. Another cycle is the long-term organic carbon cycle that contains the geological processes of sedimentation and burial of terrestrial soil and marine sediments (formation of fossil fuels coal and petroleum) to sedimentary rocks. Chemical weathering of these rocks and respiration brings CO_2 back into the atmosphere. Atmospheric CO_2 is further part of the inorganic carbon cycle, where carbon dioxide is dissolved in rain and seawater. Finally, carbon dioxide is released from the Earth's mantle through plate tectonics at ocean ridges and volcanoes, which is called the carbonate–silicate geochemical cycle.

Atmospheric carbon dioxide is a gas that has strong absorption bands in the infrared spectrum where the Earth's surface emits thermal radiation (Figures 1 and 2). Increasing absorption by CO_2 increases the atmospheric temperature and raises the effective radiative emission height. CO_2 is, hence, a potent GHG. Anthropogenic CO_2 emissions from fossil fuel burning and cement production influence these cycles. Deforestation reduces the uptake strength and thus contributes to increasing atmospheric CO_2 concentrations. About 7 giga tons of carbon is currently emitted into the atmosphere every year by these anthropogenic processes.[37] Note that only about 40% of the CO_2 emitted by humans remains in the atmosphere.[38] The oceans and the vegetation take up the majority of the anthropogenic CO_2 emissions. This means that presently the natural carbon cycles provide negative feedbacks. Because of the interactions with the organic and inorganic carbon cycles, the estimated lifetime of the emitted anthropogenic CO_2 is not easily determined. Shine et al.[4] assume an average lifetime of 1000 years in their calculations, but a significant portion can remain in the atmosphere for millennia.

Methane (CH_4)

With broad absorption bands in the infrared, methane ranks second in radiative forcing after carbon dioxide (Figure 4). It is a reactive trace gas, which is naturally produced in biological, chemical, and geological processes. In the current climate nonbiological (also called nonbiogenic), emissions are small and less than 30% compared to emissions from biological (biogenic) sources. Naturally, nonbiogenic emissions include geothermal, volcanic, and sediment emissions (CH_4 hydrates). Anthropogenic emissions are from fossil fuel mining, burning of biomass, natural gas, petroleum and coal, and waste treatment. Sources of biogenic methane include wetlands, rice agriculture, livestock, landfills, forests, termites, and also oceans. Most of the current methane in the atmosphere is anthropogenic with a rather short lifetime of about 7–10 years.[39] Biogenic CH_4 emissions from agriculture may be better controllable than CO_2 emissions; hence, reducing CH_4 concentrations has been considered as an effective alternative abatement strategy for preventing climate change.[40] Like all reactive compounds, CH_4 takes part in chemical processes with other atmospheric radicals, for example, ozone. The concentrations of methane and the other radicals are, therefore, interdependent. Methane is mainly destroyed in the stratosphere where it contributes to the production of another GHG, namely, stratospheric water vapor.[41] The climate forcing of methane can thus be considered twofold: direct, by absorption of infrared radiation, and indirect as contributor to the production of other GHGs. Methane abundance in the atmosphere increased substantially from preindustrial era to present day and has stabilized in recent years, but the causes of the stabilization is still debated.[39]

Other Kyoto Protocol Gases

Other GHGs or groups of gases listed in the Kyoto Protocol are N_2O, HFCs, PFCs, and SF_6.[42] N_2O is the fourth strongest GHG with a long lifetime of 114 years. The gas is also linked to stratospheric ozone depletion. Nitrogen fertilization and expansion of agricultural land increasingly disturb the nitrogen cycle and hence inflict a radiative forcing through N_2O. HFCs, PFCs, and SF_6 gases all stem from industrial production. The atmospheric lifetimes of these gases or groups of gases range from 1000 to 50,000 years.[42] These trace gases very effectively absorb energy in the infrared and therefore constitute potent GHGs, albeit their atmospheric concentrations are very small.[6]

Montreal Protocol Gases

In 1987, the Montreal Protocol on Substances that Deplete the Ozone Layer was designed to phase out industrial production of many chemical substances. The Montreal Protocol gases are also highly effective GHGs. This group of gases includes chloroflorocarbons, hydrochloroflorocarbons, chlorocarbons, bromocarbons, and halons. Direct air measurements show that the Montreal Protocol gases contribute about 12% to the radiative forcing of all long-lived

GHGs.[41] While successes in reducing emissions of the Montreal Protocol have been significant, the longevity of the substances between 45 and 85 years means that the actual concentrations decrease only by about 1 or 2% per year.[6,42] Note that apart from destruction by UV-light and X-rays, chemical reactions with ozone in the stratosphere are the only sinks of these inert gases.

Ozone

The reactive gas ozone is produced and destroyed by ultraviolet sunlight in photochemical reactions with other gases in the troposphere and stratosphere. Ozone concentrations vary significantly in space and time with a maximum in the tropics in about 25 km altitude. Ozone chemistry is essential for life on Earth by preventing dangerous UV sunlight from reaching the surface. Apart from absorbing UV radiation, ozone has absorption bands in the visible and infrared part of the radiative spectrum and is therefore a GHG. Environmental conditions such as wind, temperature (e.g., sudden stratospheric warming events), and irradiance play important roles in determining ozone concentrations in the stratosphere and troposphere. In the last decades, ozone increased in the troposphere, mainly due to increasing concentrations of CH_4, nitrogen oxides (NO_x), carbon monoxide (CO), and volatile organic compounds (VOCs). Ozone concentrations rather decreased in the stratosphere due to emission of man-made ozone destroying Montreal Protocol gases. Solar and volcanic forcings influence ozone concentrations in the stratosphere as well. In general, global ozone amounts decreased from the late 1970s on to around 1992 to 1993 by about 6% and are presently about 4% below the 1964 background values.[41] The overall thinner ozone layer with reduced infrared radiation causes the stratosphere and to a lesser degree the troposphere to cool. In the coming decades, the expected ozone hole recovery over Antarctica in September and October will have the opposite effect on climate and a positive, albeit regional forcing is anticipated.

Tropospheric Aerosols

In the atmosphere, solid and liquid particles in submicron size range (aerosols) reduce visibility and are commonly regarded as air pollution. Aerosol can be emitted either as particles, drops, or can be formed in the atmosphere as end products of photochemical reactions of formerly emitted natural or man-made gases (e.g., sulfur dioxide, nitrogen oxides, VOCs). The majority of aerosol mass is natural and consists of soil dust from deserts, sea salt, and dimethyl-sulfide from phytoplankton of ocean surfaces, biogenic material from plant emissions, and biomass burning aerosols from fires.[39] Parallel to GHGs, aerosol particle concentrations in the atmosphere have increased steadily with industrialization.

With regard to climate impact, the most important man-made aerosol species are sulfate, nitrate, and carbonaceous aerosol (black and organic carbon) from fossil fuel, biomass burning, and agriculture. Atmospheric aerosol layers are often complex mixes of several natural and anthropogenic particle species. The role of aerosols as climate-forcing agent has been less publicized than the role of GHGs, which may be due to local and temporally extremely variable concentrations. The average lifetime ranges from a few days to up to a week, albeit under the right circumstances, aerosol clouds can circle around the globe before they eventually fall or rain out. The direct radiative forcing of aerosols depends mainly on size and chemical composition with smaller particles scattering more effectively then larger ones.[43] This backscattering to space (mainly by sulfate) or dimming of sunlight at the surface[44] is generally a negative forcing similar to the aforementioned solar forcing. Carbonaceous aerosol (black and organic carbon) and desert dust particles additionally absorb a portion of solar or even near infrared radiation and hence constitute a positive forcing.[45,46] The emission, formation, transport, and chemical transformation of aerosols depend on local environmental conditions, in particular, wind speed, temperature, and humidity. Climate responses of other forcings that affect these meteorological parameters can feed back to aerosol concentrations and hence modify the aerosol forcing and effects. This has been the case throughout the Earth's history particularly in much dustier, while drier ice age climates.[47]

The most significant climate impacts of aerosols may be on clouds[48] and can therefore partially be considered nonradiative. If water soluble, aerosol particles serve as nuclei for cloud droplets or ice crystals where water condenses or freezes on. Sulfate particles, for example, are water soluble while soot particles are insoluble. The higher the concentration of nuclei in a volume of moist air, the more cloud droplets can form, and with limited amount of water vapor, these droplets become smaller in size. Aerosol-polluted clouds are therefore more reflective and appear darker from the surface of the Earth, which causes a negative radiative forcing. Polluted clouds may also rainout less because the average cloud droplet size and hence weight is reduced. The suppression of rain is an example of a nonradiative climate effect.

Other microphysical processes involving aerosols and clouds have been suggested in recent years.[49] Many of these indirect effects are studied

theoretically or in laboratories. Because of the complexities of cloud physics, distinguishing individual indirect effects in the field, however, is very difficult. Consequently, radiative and nonradiative impacts of aerosols are highly uncertain and global estimates depend on the few climate models capable of resolving at least some of these complexities.[50] A third category, where aerosols impact climate, is through locally or regionally modifying the energy budgets of the surface or the atmospheric layer where aerosols occur. These regions of localized cooling or heating can initiate adjustments of the circulation that can lead to modifications of weather patterns (e.g., Sahel region[51] and Indian monsoon[52]).

Over the past century, steadily increasing anthropogenic aerosol concentrations inflicted an increasingly negative forcing on the climate system that masked portions of global warming signal and likely modified rainfall.[53] Analyses of historic radiative measurements at the surface[44,54] and satellite measurements over the past decades indicated evidence of a global aerosol forcing strength of about 40% of the GHGs forcing.[55] Because of the very short lifetime of these particles, this instantaneous negative forcing can weaken quickly. Spatially resolved understanding of the evolution of the forcing is therefore needed. This volatility of aerosol forcing compared to GHGs is a major constraint of the forcing concept and the comparability of forcings.[6]

Land-use and Land Management Change

In the IPCC reports, land-use change has been included as perturbation of the reflectivity of the surface by changing vegetation types from, e.g., forest to grassland (Figure 4). Furthermore, land-use change is included in the IPCC report as an emission source of aerosols and most importantly as a CO_2 source and sink.[39] The effects of land-use change and land management in modifying the water cycle, specifically water vapor,[56] which exert additional radiative forcings, are not included in the IPCC reports. For example, trees transpire large amounts of water vapor, which has a direct effect on infrared and visible radiative absorption. Human influences, through deforestation, therefore impose disturbances in tree transpiration and hence inflict a water vapor forcing, which is especially relevant in the tropics.[57] Drying of soil is another, although nonradiative implication of deforestation. Various vegetation types or bare soil hold water differently. Thus, the evaporative cooling of the surface varies with different land coverage. These disturbances have implications for atmospheric dynamics and for convective cloud formation, which can inflict local to regional radiative forcings.[58] Changing vegetation types through land use can also disturb surface wind patterns and lead to modified emissions of trace gases and biogenic aerosols.[59] Fire management further affects the release of trace gases and aerosols, either through suppression or initiation of fires. The resulting aerosol forcings are transient and can become regionally and seasonally important.

As mentioned above in the sections of long-lived GHGs, the terrestrial carbon cycle regulates key negative feedbacks.[38] Over the course of the Anthropocene, land-use change and management may have played the second most important role in modifying atmospheric CO_2 and CH_4 amounts, after fossil fuel burning.[59] Future emission scenarios for CO_2 and CH_4 due to land-use change and management have the largest uncertainty in the global carbon budget.[60] With the latest development in Earth system modeling, it has been shown that terrestrial carbon storage capacities and biogenic aerosol production rates differ under changing climatic conditions.[39] This means that forcings from land use and management perturbations additionally depend on the climate and are, therefore, difficult to prescribe,[61] which limits the applicability of the climate-forcing concept.

Radiative Forcings Based on Economic Sectors

In the recent literature, the concept of human-activity-based radiative forcings (such as land-use change) has been embraced and estimates of radiative forcings of various economic sectors were introduced.[62] This approach considers a mixture of individual emissions of long-lived GHGs, aerosols, ozone, and other chemically active gases for each sector. Note some sectors emit species with opposing climate effects such as sulfate aerosols and GHGs at the same time. For example, Unger et al.[63] estimate the forcings of 13 economic sectors such as electric power production, household biomass burning, on-road traffic, etc. The chemical mix of the emissions of each sector is also specific to the region of origin. Hence, each sector can be characterized by a forcing with a specific temporal and spatial evolution. The advantage of this approach is the possibility of developing more mitigation oriented decision-making tools and better assessments of the short and long-term climate impacts of various sectors.

CLIMATE-FORCING DEFINITIONS, DERIVATIVES, AND ALTERNATIVE METRICS

With growing scientific insight, several attempts have been made over the years to improve the concept

of climate forcing by reevaluating its definition[6] and including more feedback processes. The most valuable definition would be one, where the global climate response to a given forcing is independent of the kind of perturbation and the climate model used. The most commonly used derivatives of the previously introduced climate-forcing definition are described below.

Instantaneous forcing (F_i)

The simplest definition of climate forcing is the 'instantaneous' radiative flux change caused by a perturbation at the top of the atmosphere or at the tropopause level (Figure 3). This flux change is calculated with a standard radiative transfer model with fixed atmospheric background conditions and with the introduction of a perturbation.[31] As illustrated in the schematic in Figure 5 on the very left, 'instantaneous' refers to the fact that the atmospheric temperature profile is not yet adjusted. With the standard GCM experiments of doubling of CO_2 and 2% solar forcing, Hansen et al.[31] showed that the resulting net radiative imbalances at the tropopause levels correlate well with the equilibrium surface temperature changes in climate models.

Adjusted Forcing (F_a)

Observations prove that the stratospheric temperature profile adjusts to a new equilibrium within months of a perturbation, whereas temperature adjustments in the lower atmosphere can take decades to centuries because of the thermal inertia of oceans and land mass. Hence, an alternative to the instantaneous forcing definition, the stratospheric 'adjusted forcing', was developed to account for the fast stratospheric adjustment (second profile pair from left in Figure 5). This forcing is calculated by allowing the temperature profile to adjust to equilibrium in the stratosphere, while the tropospheric temperature profile remains unchanged, as was the case for the instantaneous forcing. Because of the adjustment, the forcing can best be described at the tropopause level. The 'adjusted forcing' definition at tropopause height has been adapted by IPCC as the standard definition of choice since the Third Assessment Report (TAR).[3] A problem with this definition is that the tropopause height needs to be known or calculated. The tropopause height increases from pole to equator and can be set to a climatologic level of 100 hPa pressure level for a simplified forcing calculation. For climate change drivers such as tropospheric aerosols, the perturbation acts so close to the surface that the difference between adjusted forcing at the tropopause level and instantaneous forcing at the top of the atmosphere is insignificant.[64]

Fixed SST Forcing (F_s)

Atmospheric GCMs sometimes utilize as lower boundary conditions the mean state of globally measured temperatures of the ocean surface water (called sea surface temperatures or SSTs) and sea ice extent. This method[65] provides a realistic representation of the climate state where a perturbation is forced on. The resulting net radiative imbalance at the top of the atmosphere (F_0), between the control and the perturbed equilibrium run, consequently takes account of the feedbacks in the atmosphere and at the land surface but not the ocean feedbacks. Thus, F_0 is the energy flux that would ultimately be absorbed by the oceans. In this case, the atmospheric profiles and the surface conditions on land are allowed to adjust freely

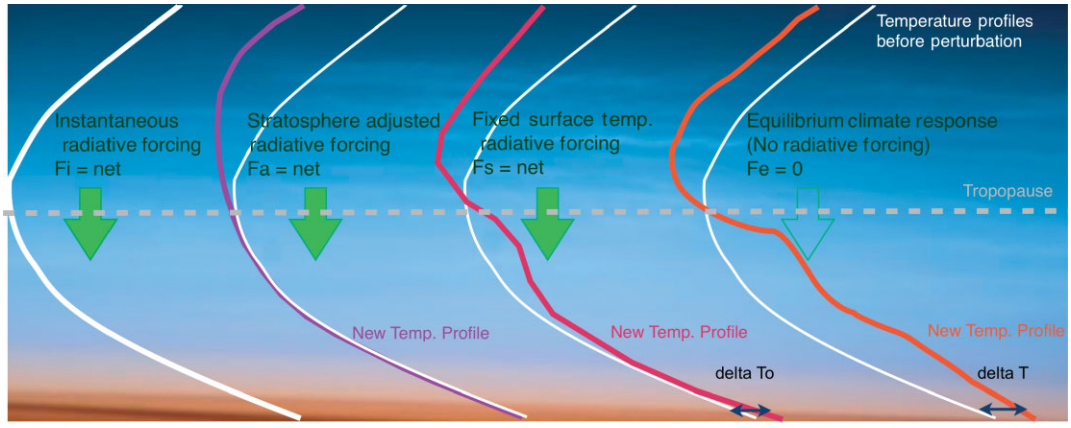

FIGURE 5 | Schematic of atmospheric temperature adjustments to radiative forcings. The conceptual framework of climate forcings and temperature profiles before (white) and after the perturbation was applied for different versions of atmosphere–ocean/land adjustments (colored lines) is shown in Figure 5. The gray dashed line illustrates the tropopause level where the various forcings are symbolized as arrows. For further information see text and also Ref 31.

to a perturbation as illustrated in the third profile pair of Figure 5. The global near surface temperature change ΔT will need to be added to the 'fixed SST forcing' accordingly:

$$F_0 + \frac{1}{\lambda} \cdot \Delta T_0 = F_S. \quad (2)$$

This definition is particularly useful for forcings that are nonradiative in the first place. For these cases, atmospheric or land adjustments are necessary to create realistic radiative imbalances that do not underestimate the impacts of these climate change drivers. The top of the atmosphere imbalance F_0 is sometimes referred to as 'radiative flux perturbation'.[66] Tropospheric aerosols and land-use change are important examples, where including the land and atmosphere adjustments is crucial.

Climate Response

The right profile pair in Figure 5 symbolizes the fully adjusted climate response to a perturbation, together with the unperturbed temperature profile. In the fully adjusted profile, the radiative fluxes are more or less in equilibrium everywhere. Hence, the radiative forcing at the tropopause is shown as zero. In this case, Gregory et al.[67] suggest that climate forcings can be estimated from the change of the global mean surface temperature (ΔT_0) and the change in net radiative flux at the top of the atmosphere (F_0), while the modeled climate system is in the process of adjusting to equilibrium. Naturally, F_0 is zero when the system reaches steady state. The forcing F_S can then be estimated from the intercept, in a regression plot of Eq. (2). Technically, this estimate is not a forcing anymore because all atmospheric and feedback mechanisms are in place. Therefore, the calculated 'forcing' depends strongly on the specifics of the climate model experiment.[68,69] On the other hand, this method does not require extra radiative model runs or equilibrium runs.

Global Temperature Change

The expected global temperature change for a given perturbation, after the system reaches equilibrium, may be the most descriptive measure for climate impact (see ΔT in Figure 5 right profiles). It is increasingly used as benchmark for future vulnerability studies.[5,69] On the other hand, calculating global temperature changes for a given perturbation requires significant GCM modeling capabilities and not just radiative calculations. Furthermore, the results will significantly depend on model performance, which impedes model-independent comparisons of perturbations.

Efficacy

The main assumption of the forcing concept is that climate response is similar for a wide range of forcing agents. This assumption enables comparisons of climate impacts of various Kyoto-regulated GHGs or groups of gases ('comprehensiveness and cost effectiveness'). However, in some cases (i.e., anthropogenic aerosols and ozone), the linearity and comparability becomes less clear. Note that some impacts are unique to the specific spatial distribution of a forcing agent. To ensure compatibility of climate forcings for various agents in spite of these shortcomings, scientists introduced the 'efficacy' of climate forcing.[65,70] It is defined as the global mean temperature change per unit forcing produced by a perturbation, relative to the response produced by a standard CO_2 forcing at the same initial state. Hansen et al.[65] showed that a broad variety of efficacies exist. For example, anthropogenic methane has efficacy of 110%, which increases to 145% when its indirect effects on stratospheric water vapor and tropospheric ozone is included. Tropospheric aerosol efficacy also differs significantly from unity.[65]

ENERGY BUDGET APPROACHES

Based on the explanations above, it can be argued that the more complex the interactions between perturbations and the climate system, the radiative forcing concept becomes the less applicable. To circumvent these shortcomings, alternative approaches have been suggested. Investigations of the full energy budget (radiative plus nonradiative fluxes) promise physically pure alternatives, albeit their executions may be modeling intensive. The energy budget approach can be considered at the top of the atmosphere or at the surface, depending on the scientific question.

Ocean Heat Content

The climate system has considerable thermal inertia due to the ocean heat capacity, which is so large that additional heat can be absorbed and stored for centuries. Therefore, a radiative imbalance at the top of the atmosphere can persist over decades or centuries after the initial forcing was introduced. The ocean–atmosphere system thereby returns only slowly back to balance. Hansen et al.[71] estimate that with a climate sensitivity λ of 0.75 ± 0.25 K per W/m^2 the climate system will need 25–50 years for the global surface temperature to reach 60% of its

equilibrium response. In the context of anthropogenic changes, improvements in precise measurements of ocean heat fluxes over the last decades provide independent estimates of the radiative imbalances of the system.[72–74] Hansen et al.[71] infer from these data that in 2003 the Earth was out of balance by 0.85 ± 0.15 W/m^2. This means an additional global warming of about 0.6 K is still 'in the pipeline' (meaning stored in the oceans) without adding any further perturbation.

Top of the Atmosphere Net Radiative Flux

Individual forcings cannot be measured directly, although satellites have been measuring the adjusted net radiative imbalance (F_0 in Eq. (2)) at the top of the atmosphere for a few decades.[14] On the basis of these observations and radiative transfer calculations, Murphy et al.[55] estimate the energy imbalance of the Earth from 1950 to present. According to this study, only about 10% of the GHG and solar forcing has been used to heat the climate system. The oceans mainly absorbed this energy and about 20% of it was emitted back to space. Volcanic aerosols further balanced about 20% of the forcing by reflecting shortwave radiation back to space. The remaining 50% of the forcing could have been contributed by increasing concentrations of tropospheric aerosols, their direct forcings, and their indirect effects, in spite of their strong spatial and short-term variability.

Surface Energy Budget

The formulation that links together all the energy transfer processes of different time scales in the atmosphere–ocean/land system is the global surface energy budget. The changes in net radiative flux balance the changes in turbulent heat release due to convective and evaporative cooling, changes in melting of snow, ice, and ocean heat uptake. The surface fluxes can be measured quite accurately and can also be predicted with GCMs. Furthermore, instantaneous and adjusted radiative forcings can in principle be estimated with GCM experiments at the surface as well.[68] However, these surface forcings may differ significantly in size from classic forcings. For example, the instantaneous forcing of absorbing aerosols (the so-called Asian Brown Clouds[46,75]) can be much smaller at the top of the atmosphere compared to the surface. Furthermore, in the context of climate change, the main advantage of considering the surface energy budget lies in the link to the water cycle.[53,76] Specifically, the surface energy budget constrains the energy used for evaporation. Global mean evaporation changes are equal to changes in precipitation because the water holding capacity of the atmosphere is small.[77] With equilibrium GCM experiments, Liepert et al.[53] argue that the surface energy budget approach is important for determining global precipitation changes particularly when forcings are spatially inhomogeneous and interact with each other, such as tropospheric aerosols and GHGs.

CONCLUSION

The concept of climate forcing relates perturbations in the climate system to climate responses (namely, global warming) through their feedbacks and climate sensitivity. Climate forcings, hence, quantify the strengths of various perturbations in terms of radiative imbalances at the tropopause or the top of the atmosphere. Thereby, they solve the problem of comparability of various climate change drivers. This request was amended by the UNFCCC for assessing the cost effectiveness and comprehensiveness of abatement strategies and political and economic adaptations to climate change. The concept of climate forcing, which originally stems from paleo-climate and climate modeling studies, should ideally have the following properties:

- independence of climate response from type of climate-forcing agent;
- additive character of forcings;
- independence of climate forcing from type of climate model used to calculate responses; and
- convenience of calculation.

The IPCC chose as definition, the adjusted forcing at the tropopause level, which comes closest to fulfilling the desired properties, albeit limitations exist and have been summarized here. For example, climate response is not always independent of the type of forcing and their geographic distribution. Absorbing aerosols have a significantly lower surface temperature response compared to GHGs for a unit of forcing. Hansen et al.[65] address this issue by adding appropriate weights, also called 'efficacies', to the forcings. Another example is ozone forcing, which violates the additive character of forcings. Ozone chemistry affects the strengths of other GHG forcings, which need to be corrected accordingly. Furthermore, the calculated forcings loose their model independence if clouds and processes at the land surfaces are involved in the perturbations. Land-use change and aerosol indirect effects are such forcings that disturb clouds and rainfall in nonradiative ways, which cannot properly be addressed by the forcing concept. When nonradiative processes are

included in the forcing calculation, the result will become dependent on model performance. Recently, methods have been developed, and are described here, that focus on indirect evaluations of the strengths of forcings with observations. In these approaches, observations of ocean heat uptake, satellite records of radiative imbalances, and surface energy budgets are combined with radiative calculations. Interesting new ideas are published that target abatement strategies more directly by calculating the forcing of classes of polluters (i.e., heavy traffic) through bundling emissions of various trace gases and aerosols. Future research will need to expand the forcing concept by adding the water and the carbon cycles. Including precipitation as a climate response will also need to be addressed. New avenues are surface energy budget approaches that link the water cycle with the radiative forcing concept. A major problem of these approaches, however, is the dependence on climate modeling for predicting climate responses. This issue can be addressed by focusing the attention on fewer but well-constructed model experiments and careful evaluation of the underlying physical, chemical, and biological basics. Alternatives to the forcing concept can be developed with constrained models and thorough analysis of the climate processes in observations.

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