

## Chapter 2

# Variability and Change in Climate

**Abstract** Understanding of the current and future climate change requires understanding of mechanisms which controlled climate change both before and after the last glacier ages. Although the Sun played an important role in climate variability and climate change in paleoclimate, it is generally accepted that the Sun has not been a major driver of climate change since the Industrial Era. This chapter describes the observed climate change with particular emphasis on the Industrial Era period. The globally averaged atmospheric mole fraction of carbon dioxide (CO<sub>2</sub>) reached the abundance of 144% relative to the preindustrial concentrations in 2015, with many other climate variables setting new records in the past few years. For example, atmospheric abundance of methane (CH<sub>4</sub>) and nitrous oxide (N<sub>2</sub>O) reached  $1845 \pm 2$  and  $328 \pm 0.1$  ppb in 2015, respectively. Also, 5 major and 15 minor greenhouse gases (GHGs) contributed  $2.94 \text{ W m}^{-2}$  of the direct radiative forcing which is 36% greater than their contribution at the onset of Industrial Revolution in 1750. The record high radiative forcing has resulted in the highest annual global surface temperature over  $\sim 135$  years of modern record keeping. Since oceans absorb about a quarter of anthropogenic CO<sub>2</sub> emissions, increase in CO<sub>2</sub> concentration during the Industrial Era has resulted in ocean acidification equivalent to approximately 30% increase in H<sup>+</sup> concentration in ocean water. Other associated changes include increase in sea surface temperature (SST) and increased thermal energy content of the ocean which absorbs 90% of Earths' excess heat from GHG forcing. Also ocean warming and increased stratification of the upper ocean caused by global climate change results in deoxygenation of interior oceans with implications for ocean productivity, nutrient cycling, C cycling and marine habitat. Owing to ocean warming and ice melting, the global sea level rise reached 67 mm greater than the 1993 annual mean, when satellite altimetry measurement began, with salty regions of ocean getting saltier while fresh water regions of ocean parts are getting fresher. CO<sub>2</sub> therefore, remains the single most important anthropogenic GHG, contributing 65% to long-lived GHGs radiative forcing, and responsible for nearly 83% of the increase in radiative forcing over the decade ending in 2015.

**Keywords** Climate • Global warming • Greenhouse gases • Solar radiation • Solar activity • Ecosystems

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## 2.1 Introduction

The Earth’s climate system is powered by solar electromagnetic radiation, and any variability of the Sun’s radiative output has the potential of affecting climate and, hence the habitability of the Earth (Solanki et al. 2013). However, changes in solar power output on decadal, centennial, and millennial timescales are limited to small changes in the effective global surface temperature on shorter timescales (Lockwood 2012). Sun is the source of practically all external energy input into the climate system. Change in solar radiation can influence climate system through: (a) variations in insolation caused by changes in the Sun’s radiative output (i.e., direct influence), (b) modulations of radiation reaching different hemispheres of the Earth through changes in Earth’s orbital parameters and in the obliquity of its rotation axis (i.e., indirect influence through changes in Earth’s orbit), (c) alterations in the influence of Sun’s activity on Galactic cosmic rays which affect the cloud cover (e.g., Marsh and Svensmark 2000), (d) alterations of the fraction of solar radiation that is reflected (a fraction called albedo—it can be changed, for example, by changes in cloud cover, aerosols, or land cover), and (e) variations in the long wave energy radiated back to space (e.g., through changes in atmospheric GHG concentration). In addition, local climate also depends on distribution of heat by winds and ocean currents. All these factors have played some role in the past climate change.

The first of these is generally considered to be the main cause of the solar contribution to the global climate change. Solar variability takes place at many time scales that include 27-day variation due to solar rotation, annual variation due to ellipticity of Earth’s orbit, decadal scale solar magnetic cycle (sunspot), and also the oscillation between grand solar maxima and minima on timescales of several centuries (Helama et al. 2010; Lockwood 2012). Variability associated with the 11-year solar cycle has shown to produce measurable short-term climate anomaly (Gray et al. 2010; Lockwood 2012). There is a growing evidence suggesting that changes in solar irradiance affect Earth’s middle (between 10 and 50 km) and lower atmosphere (Gray et al. 2010). The second is believed to be the prime cause of the patterns of

glacial-interglacial cycles of Ice Age that have dominated the longer term evolution of the climate over the past few million years (refer Chap. 5). The various parameters of Earth's orbital and rotational motion vary at periods of 23,000 years (precession), 41,000 years (obliquity), and 100,000 years (eccentricity) (Crucifix et al. 2006). The third path builds on the modulation of the Galactic cosmic rays by solar magnetic activity. The Sun's open magnetic flux and the solar wind impede the propagation of the charged Galactic cosmic rays into the inner Solar System, so that at times of high solar activity, fewer cosmic rays reach the Earth. Their connection with climate has been drawn from the correlation between the cosmic rays flux and global cloud cover (Marsh and Svensmark 2000). Most of solar irradiance variations are produced by dark (sunspots) and bright (magnetic elements forming faculae) surface magnetic structures on the solar surface, whose concentration changes over the solar cycle. Variations in solar irradiance produce natural forcing of Earth's climate with global and regional scale responses (Lean and Rind 2008). Globally the mean surface temperature varies in phase with solar activity. A number of attempts at reconstructing past climate variations through solar forcing have indicated that before 1940, direct forcing by changes in solar irradiance can explain the large proportion of observed changes. However, it is virtually impossible to assign the global warming of the past half century, and particularly the observed warming since 1975 to variations in solar irradiance alone using either statistical or physical methods. The indirect solar forcing seems to be small and unable to explain the observed warming (Keller 2009).

Earth's climate is the result of balance between incident shortwave solar radiation absorbed and long wave infrared radiation emitted (described in Chap. 1). As the Earth's temperature has remained relatively constant over many centuries, the incoming solar energy must have remained nearly in balance with outgoing radiation. A net change in imposed perturbation of this radiation balance to Earth system, either through anthropogenic activity or natural process is referred to as a *radiative forcing*. The radiative forcing averaged over a particular length of time quantifies the energy imbalance that occurs when the imposed change takes place. Changes in the atmosphere, land, ocean, biosphere, and cryosphere; either natural or anthropogenic can perturb the Earth's radiation budget and produce radiative forcing that could affect the climate.

Changes in global energy budget occur either from changes in net incoming solar radiation or changes in outgoing long wave radiation. Changes in net incoming solar radiation derive from changes in the net Sun's output of energy. A precise record of measurements of total solar irradiance (i.e., the total power of Sun affecting a unit area perpendicular to the Sun's rays in  $\text{W m}^{-2}$ ) extending back to 1978 established a generally accepted value of about  $1360.8 \pm 0.5 \text{ W m}^{-2}$  (Kopp and Lean 2011; Kopp et al. 2012; Solanki et al. 2013), which is responsible for keeping Earth from cooling off to temperatures that are too low for sustaining human life. The composition, structure, and dynamics of Earth's atmosphere also play a fundamental role of making efficient use of the energy input from the sun

through the greenhouse effect. The net imbalance between absorbed shortwave radiation and outgoing long wave radiation at the top of the Earth's atmosphere is a fundamental climate variable which represent a nexus between changes in radiative forcing that sets the trajectory of climate change and climate response. The magnitude of climate response is determined by feedbacks which may amplify or diminish climate responses, but also influenced by unforced variability internal to climate system (Hansen et al. 2011). Surface energy fluxes drive ocean circulation, determine how much water is evaporated from the Earth's surface, and govern the planetary hydrological cycle.

Changes in outgoing long wave radiation can result from changes in temperature of Earth's surface or atmosphere or changes in emissivity (i.e., measure of emission efficiency) of long wave radiation or from atmosphere or Earth's surface. Ocean heat content and satellite measurements indicate a small positive energy imbalance (Trenberth et al. 2009, 2015; Murphy et al. 2009; Stephens and L'Ecuyer 2015) that is consistent with rapid changes in atmospheric composition. Over the 1985–1999 period, the imbalance between absorbed shortwave radiation and outgoing long wave radiation at the top of atmosphere was  $0.34 \pm 0.67 \text{ W m}^{-2}$ , and increased to  $0.62 \pm 0.43 \text{ W m}^{-2}$  for 2000–2012 period, despite slower rate of surface temperature increase since 2000 compared with late 20th century (Allan et al. 2014). The changing atmospheric composition, especially due to increase in atmospheric GHGs from human activities, including burning of fossil fuels is responsible for the observed imbalance. Increasing GHGs in the atmosphere causes imbalance to inflow and outflow of energy to the Earth system at the top of the atmosphere by increasingly trapping more radiation, and therefore, creating warming (Trenberth 2009), which can be manifested in many ways, including rising surface temperature, melting Arctic sea ice, increasing the water cycle, and altering storms. Most of excess energy goes into the ocean, however (Trenberth 2009; Bindoff and Willebrand 2007). Over the past 50 years, the oceans have observed about 90% of total heat added to the climate system and the rest about 10% is used to melt sea and land ice, warming the land surface, and warming and moistening the atmosphere (Trenberth 2009). Strengthened ocean heat uptake, especially below 700 m depth after 2000 is responsible for slowing global mean surface air temperature increase during the first decade of 21st century (Watanabe et al. 2013; Balmaseda et al. 2013).

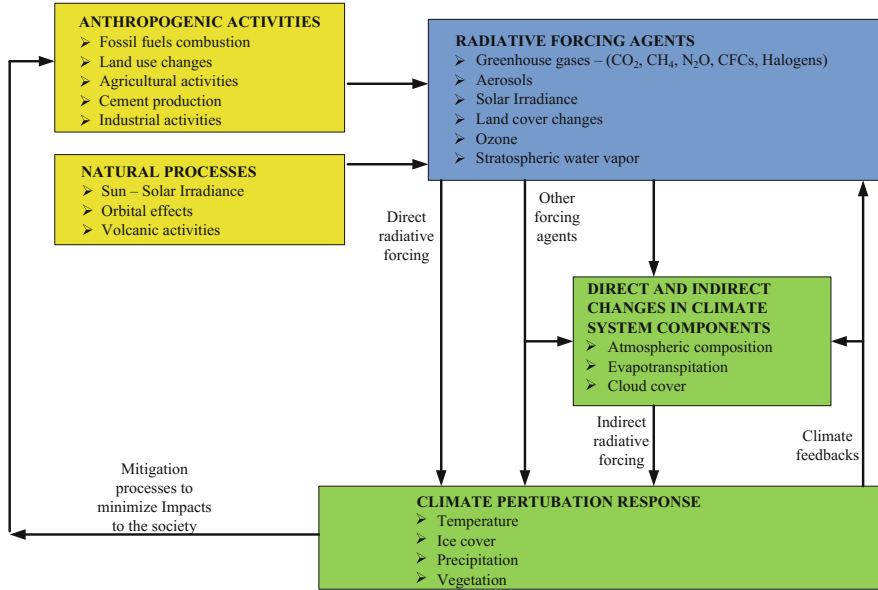
Climate change is driven by disturbances to the energy balance of the Earth system, which are generally termed climate forcings. Climate system also exhibits unforced (i.e., chaotic) variability. However, it is now widely agreed that the strong global warming trend since the late 19th century is caused predominantly by man-made changes in atmospheric composition (Hegerl and Zwiers 2007; IPCC 2014). Increase in atmospheric GHGs concentration (e.g.,  $\text{CO}_2$ ,  $\text{CH}_4$ , and  $\text{N}_2\text{O}$ ) makes the atmosphere more opaque at the infrared wavelengths, and this increased opacity reduces transmission of heat to space. The temporary imbalance between the energy absorbed from the sun and heat transmission to space causes the planet

to warm until the planetary energy equilibrium is restored. The planetary energy imbalance caused by a change of atmospheric composition defines a climate forcing. The eventual global temperature change per unit forcing (i.e., climate sensitivity) is known with reliable accuracy from Earth's paleoclimate history (Hansen et al. 2011). However, presence of aerosols such as dust, sulfates, and black soot (Ramanathan et al. 2001) in the atmosphere which can both reflect solar radiation to space (cooling effect) and absorb solar radiation (warming effect), and also the efficiency of heat mixing into the deeper ocean limit the ability to predict the global temperature on decadal timescales (Hansen et al. 2011). Ocean heat mixing is a complex and more difficult to simulate by climate models.

Changes in atmosphere, land, and cryosphere can perturb the Earth radiation budget and produce radiative forcing that affect climate. Climate feedbacks, the physical processes that comes into play as climate changes in response to forcing can either amplify or diminish the effects of change in climate forcing. Positive feedbacks amplify while negative feedbacks diminish the climate response. Climate feedbacks do not come into play coincident with the forcing, but rather in response to climate change. Feedbacks operate by altering the amount of solar energy absorbed by the planet or the amount of heat radiated into space, and they tend to be a function of global temperature change. Climate feedbacks can be grouped into fast and slow feedbacks. Fast feedbacks appear almost immediately in response to global temperature change. These include water vapor when its atmospheric concentration is enhanced by increase in surface temperatures (Hansen et al. 2011). Water vapor is a powerful GHG, and its increasing atmospheric concentration enhances the greenhouse effect and leads to further warming (Cubasch et al. 2013). Other fast feedbacks include clouds, natural aerosols, snow cover, and sea ice. Slow feedbacks may lag global temperature change by decades, centuries, millennia or longer timescales (Hansen et al. 2011). The principal slow feedbacks are changes in continental ice sheet area which affect surface reflectivity or albedo, and long-lived GHGs. The objectives of this chapter are to summarize the current knowledge on climate change and roles of both natural processes and anthropogenic activity to set the stage for discussion on the role of C cycling and climate change in the next sections.

## 2.2 Radiative Forcing

For convenience, factors responsible for climate change are generally separated into forcings and feedbacks. Forcings are energy imbalances imposed on the climate system externally by both natural processes or as a result of human activities (Fig. 2.1). Increase in atmospheric GHGs concentration, particularly  $\text{CO}_2$  and  $\text{CH}_4$  are the main anthropogenic forcing. Increase in atmospheric GHGs concentration causes warming through their greenhouse effect. As a result of warming, the surface restores the radiative balance by increasing radiation to space, but also warming causes atmospheric water vapor, albedo, clouds, vegetation, ice sheets, permafrost,



**Fig. 2.1** Anthropogenic activities and natural processes effects on climate forcing and associated climate response and feedbacks

and atmospheric chemistry to change. These changes affect the Earth’s radiation budget directly or indirectly (Fig. 2.1). Feedbacks are the internal processes that amplifies or dampens the climate response. For example, warmer global temperatures increase atmospheric water vapor, which amplifies the initial warming due to increased atmospheric concentration of GHGs through greenhouse properties of atmospheric water vapor. Also, warmer temperatures lead to melting of snow and ice, which exposes darker surface that absorbs rather than reflecting incoming solar radiation, leading to more warming and melting than it would have occurred if the snow cover had been fixed. Feedbacks can occur at a broad range of timescale, from instantaneous up to thousands of years (Wolff et al. 2015).

The Intergovernmental Panel on Climate Change (IPCC) uses radiative forcing (RF) to assess and compare externally imposed perturbation in radiative energy budget on Earth’s climate. Such perturbations can be a result of natural or anthropogenic causes or both. The RF is defined as change in net downward minus upward radiative energy flux at the top of atmosphere (tropopause) due to change in external driver of climate change (Myhre et al. 2013). The RF quantifies the perturbation in radiative fluxes caused by changes in forcing agents such as atmospheric GHGs and expressed in  $W m^{-2}$  averaged over a particular period of time. It therefore, quantifies energy imbalance in terms of temperature change that occurs when imposed change takes place. It is computed with all tropospheric properties held at their unperturbed values and allowing perturbed stratospheric temperature to readjust to radiative energy dynamic equilibrium. The IPCC analyses uses 1750

(the beginning of Industrial Era) as a benchmark for assessing changes in climate system caused by human activities and expressed as the changes due to anthropogenic activity. From the beginning of the Industrial Era, the Earth has undergone a very fast and unusual change in RF resulting from anthropogenic actions, which includes increase in atmospheric GHGs concentration, changes in concentration of aerosol particles in the atmosphere and ozone destroying chemicals in the stratosphere as well as changes in the nature of land surface (Table 2.1). The current trajectory of these changes suggests a substantial changes in climate by the end of the 21st century (IPCC 2013).

The utility of the RF concept is that it enables the quantification of various factors that shift the energy balance and assess their relative importance to climate change. The RF can be related through a linear relationship to the global mean equilibrium temperature change at the surface [Eq. 2.1]:

$$\Delta T = \lambda RF \quad (2.1)$$

where,  $\lambda$  is the climate sensitivity parameter (Ramaswamy et al. 2001). The  $\lambda$  derived with respect to RF can vary substantially across different forcing agents (Table 2.1). The forcing factors are external to climate system, and not part of it. The important forcing factors and their estimated values in 2014 are presented in Table 2.1. Equation 2.1 suggests a straightforward calculation of the equilibrium change in temperature at a global scale arising from a particular change in RF if the Earth behaved as simple black body with no additional effects (Knutti and Hegerl 2008). For example, doubling of atmospheric CO<sub>2</sub> concentration results in long-wave forcing of  $\sim 3.7 \text{ W m}^{-2}$ , which would cause an equilibrium warming of  $\sim 1.2 \text{ }^\circ\text{C}$  (Knutti and Hegerl 2008). However, in practice the initial perturbation causes a range of other feedback effects (Fig. 2.1), which may weaken (negative feedbacks) or strengthen (positive feedbacks) the global temperature response, and it is the net effect of such feedbacks that determines the sensitivity of climate to forcing. The negative feedbacks (e.g., changes in vertical temperature gradient of the atmosphere) causes the climate to be less sensitive to changes in forcing, while positive feedbacks (i.e., albedo feedbacks) causes the climate to be more sensitive. The nature and magnitude of these feedbacks are the principal cause of the uncertainty in response of global climate to different emission scenario and GHGs concentration pathway over multi-decadal and longer periods (Wolff et al. 2015). Feedbacks also play role in inducing regionally variable responses to climate forcing, both in temperature and other variables such as rainfall and occurrence of extreme events. The RF and their responses are assumed to be additive, making it a useful tool for designing policies towards a climate change mitigation target. Analysis of forcing due to observed or modeled concentration changes between pre-industrial and a selected later year provides indication of relative importance of different forcing agents during the period (Table 2.1). The combined effects of all feedbacks are significantly positive (IPCC 2013).

Application of RF concept in climate change detection and prediction has some limitations, including its inability to include other associated climate change

**Table 2.1** Global average anthropogenic and natural radiative forcing estimates for the period 1750 to 2014

Parameter	Concentration in 2015*	Lifetime (years)	Average RF 1750 to 2015 ( $\text{W m}^{-2}$ )	Range ( $\text{W m}^{-2}$ )	Source
Carbon dioxide ( $\text{CO}_2$ )	$401.6 \pm 0.1$		1.91	1.71–2.10	Butler and Montzka (2015) Myhre et al. (2013)
Methane ( $\text{CH}_4$ )	$1833 \pm 1$	9.1	0.50	0.45–0.55	
Nitrous oxide ( $\text{N}_2\text{O}$ )	$327 \pm 0.1$	131	0.19	0.16–0.22	
Well-mixed GHGs			2.60	2.32–2.87	
CFC-11	$233.5 \pm 0.8$	45	0.058		Butler and Montzka (2015) NOAA/ESRL (2016) Myhre et al. (2013)
CFC-12	$578.5 \pm 1$	100	0.166		
CFC-113	72.2	85	0.116		
HFC-125	15.95	28.2			
HFC-134a	60.7	13.2			
HFC-143a	15.31	47.1			
HFC-152a	6.63	1.5			
HFC-23	9.03	222			
HCFC-22	231.1	11.9			
HCFC-141b	24.1	9.2			
HCFC-142b	219	17.2			
$\text{CCl}_4$	83.4	26			
$\text{CH}_3\text{Cl}_3$	3.4	5			
$\text{SF}_6$	8.76	3200			
$\text{CF}_4$		50,000			
$\text{C}_2\text{F}_6$		10,000			
Total halogens	–	–		0.36	0.32–0.40
Tropospheric ozone	–	–	0.40	0.20–0.60	Myhre et al. (2013)
Stratospheric ozone		–	–0.05	–0.15 to +0.05	
Stratospheric water vapor			0.07	0.02–0.12	
Aerosols			–0.80	–1.2 to +0.15	
Surface albedo			0.04	0.02–0.09	
Contrails			0.06	0.005–0.15	
Solar irradiance			0.05	0.0 to +0.10	
Total RF	–	–	$2.94 \pm 0.03$		

\* $\text{CO}_2$  concentration in parts per million (ppm),  $\text{CH}_4$ ,  $\text{N}_2\text{O}$  concentration in parts per billion (ppb), All halogens concentration in parts per trillion (ppt)



impacts such as changes in precipitation, surface sunlight available for photosynthesis, extreme events, and regional temperatures which can differ greatly from the global mean temperature. Although it is useful in understanding global mean temperature change, it provides only limited perspective on factors driving broader climate change. In addition, the RF matrix does not allow comparison on effects such as the influence of land cover change on evapotranspiration (Andrews et al. 2012). Despite these limitations, the observed changes in climatic variables approximately scale with temperature (Tebaldi and Arblaster 2014; Herger et al. 2015), suggesting that the global temperature is probably the best proxy for the aggregated impacts of forcings and feedbacks, even though the relation is likely nonlinear. Global temperature is relatively easier to measure, and its record extends further back than measurements of most other climate variables. In addition, global temperature can be reconstructed from paleodata, which is not the case for other climate quantities.

### 2.3 Detection and Attribution of Climate Change

In the IPCC assessments, detection and attribution of climate change involves quantifying the evidence linking external drivers of climate change and observed change in climatic variables. Detection and attribution, therefore, attempts to separate the observed climate changes into components that can be explained by variability generated internally within the climate system and components that are the result of forcings external to the climate system. Atmospheric processes that generate internal climate variability are known to operate on time scales ranging from instantaneous to years. Examples of internal climate variability include water vapor condensation in clouds, inter-hemispheric exchange, and troposphere-stratosphere exchange which operate on short timescales. In addition, internal variability can also be produced by interaction between components, such as El Niño Southern Oscillation (ENSO) produced by coupled ocean-atmosphere phenomenon occurring in the tropical Pacific (Toniazzo 2006; Toniazzo and Scaife 2006; Hertig et al. 2015). Other components of climate system such as the ocean and large ice sheets operate at a longer timescales. Greenland and Arctic ice sheets are important cryosphere elements affecting both regional and global climate by causing polar amplification of surface temperatures, a source of fresh water to the ocean, and also representing a potentially irreversible change to the state of Earth system as they disappear (Jacob et al. 2012; Seo et al. 2015).

The initial goal of the detection and attribution was to determine whether RF due to GHGs increase has influenced the climate by quantifying uncertainty through simulations of temperature changes with observations (Allen et al. 2000; Houghton et al. 1996; Gillett et al. 2002; Stott and Kettleborough 2002). Subsequently, detection and attribution methods are also used to evaluate the ability of climate models to simulate the observed climate change, assess the role of external factors versus climate variability in observed climate change, and enable the prediction of

future climate change based on the changes that have been observed so far (Hegerl and Zwiers 2011; Bindoff et al. 2013; Stott et al. 2016).

Detection of change is a process of demonstrating that climate and/or the system affected by the climate has indeed changed in some defined statistical sense without providing reasons for the detected change (Hegerl et al. 2010). The identifiable change is detected in observations if the likelihood of occurrence of change by chance due to internal climate variability is small. Attribution is a process of evaluating the relative contributions of multiple causal factors responsible for the detected change or event, with an assignment of statistical confidence. Attribution requires the detection of a change in the observed variable or closely associated variables (Hegerl et al. 2010; Bindoff et al. 2013). Therefore, detection and attribution seeks to determine whether climate is changing significantly and if so, what has caused such changes. Such an understanding has several applications: (i) to know if GHGs emission are contributing significantly to climate change, and therefore need to reduce emissions if they are, (ii) to understand the current risks of extreme weather events. Under a non-stationary climate, the traditional definition of climate as statistics of the weather over a fixed 30-year period can no longer hold, since the definition assumes that the climate is stable, as had been traditionally, and what were previously rare events could be already much more common, consequently, general circulation models are needed to characterize the current climate, which can be different from that of previous or succeeding years, and (iii) by comparing observations with models prediction in a rigorous and quantitative way, attribution can improve confidence in model predictions and point out areas where models are deficient and needing improvements.

Although the observational record show clear signs of warming climate, the record does not clearly indicate the causes of the observed changes. Attribution of climate change, i.e., the process of establishing the most likely causes for the detected change with some level of confidence—seek to determine which external factors have significantly affected the climate. External forcing factors are the agents outside the climate system that cause the climate to change by altering the radiative balance or other properties of climate system. Examples of anthropogenic external forcing factors include increase in well-mixed GHGs and changes in sulfate aerosols which affect clouds and make them more reflective and scatter more incoming solar radiation to space, while the external natural forcing factors include solar radiation variability and volcanic activity. Due to internal variability, the attribution statements can never be made with 100% confidence.

The global mean temperature change that results in response to sustained perturbation on the Earth's energy balance after the allowing of enough time for the atmosphere and the oceans to achieve thermal equilibrium is termed as Earth's climate sensitivity. Climate sensitivity has the units of temperature change in  $W\ m^{-2}$ . The sensitivity of climate system to external forcing is governed by the energy imbalances they induce and partitioning of these imbalances between atmosphere, ocean and cryosphere (Trenberth et al. 2009, 2014). However, most of excess energy goes into the global oceans, and oceans act as a large heat sink (Church et al. 2011; Knutti and Rogelj 2015). Equilibrium climate sensitivity

combines changes resulting from RF and feedbacks to characterize the temperature response of the Earth to change in forcing (Knutti and Rugenstein 2015). It is defined as the equilibrium global average surface warming in response to RF from an atmospheric CO<sub>2</sub> doubling. It includes feedbacks such as the changes in water vapor, lapse rate, surface albedo and clouds (Knutti and Rugenstein 2015). It is a convenient tool in modeling and policy making for emission control. The incoming solar radiation can also be affected by natural forcing factors including changes in output from the Sun and changes in stratospheric aerosols resulting from volcanic eruptions. The effect of solar forcing on global mean surface temperature trends is considered small, however, with less than 0.1 °C warming attributable to combined solar and volcanic forcing over the period 1951–2010 (Jones et al. 2012a). Variabilities associated with 11-year solar cycle produce measurable short-term regional and seasonal climate anomalies (Lockwood 2012; Gao et al. 2015).

## 2.4 Climate Change

The Earth's climate history in the past one million years has varied from cold 'icehouse' conditions, with a documented cold climate and a sequence of glacial-interglacial cycles (Augustin et al. 2004), to a 'warmhouse' conditions when glaciers generally disappeared in the Northern Hemisphere. The onset of melting of the last glacial maximum ice sheets occurred at approximately 20,000 years ago, a period generally termed as 'last glacial termination'. It was followed by warmer 'Holocene', the current interglacial period, where the climate has remained warmer and remarkably stable compared to glacial-interglacial period and favorable for human civilization to flourish. During this stable climate, there have been notable regional climatic fluctuations, of which, the most notable include the period known as 'Little Ice Age' from 1600 to 1800 when Europe experienced unusually cold conditions and expanded state of glaciers globally (Mann 2002; Matthews and Briffa 2005). Since industrial revolution in 1750, however, increasing evidence points to large human impacts on the planet and global climate to the extent that some scientists and scholars have termed this period as 'Anthropocene' suggesting that human beings have overwhelmed the forces of nature and become the dominant drivers of global change (Steffen et al. 2007; Waters et al. 2016).

The IPCC (2013) defines climate change as a change in the state of the climate that can be identified using statistical test by changes in mean and/or variability of its properties, and persists for an extended period, typically decades or longer. These changes can be due to either natural processes and/or external forcings. Some external influences such as changes in solar radiation or volcanic activities occur naturally, and can perturb radiation budget while causing variability of the climate system. However, its estimated contribution to currently observed climate change is small (Bindoff et al. 2013). Volcanic eruptions inject aerosols to altitudes as high as 10–30 km in the stratosphere, where they reside for 1–2 years, reflecting sunlight and cooling Earth's surface (Hansen et al. 2011).

The drivers of changes in climate, therefore, include: (i) solar irradiance, (ii) aerosols, (iii) clouds, (iv) ozone, (v) surface albedo changes, and (vi) changes in atmospheric GHGs concentration. The principal global anthropogenic forcing are GHGs and the tropospheric aerosols, mostly in the lower few kilometers of the atmosphere (Myhre et al. 2013). Well-mixed GHGs (e.g., CO<sub>2</sub>, CH<sub>4</sub>, and N<sub>2</sub>O) are closely linked to anthropogenic activities and they also interact strongly with the biosphere and the oceans. Their earlier atmospheric histories have been reconstructed from measurements of air stored in archives trapped in polar ice cores or in firn, and established the pre-industrial (1750) mole fraction of  $278 \pm 2$  ppm,  $722 \pm 25$  ppb, and  $270 \pm 7$  ppb for CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O, respectively (Etheridge et al. 1996, 1998; Prather et al. 2012). Anthropogenic activity increases the atmospheric concentrations of well mixed GHGs, aerosols, and cloudiness. The principal GHGs emitted from anthropogenic activity during industrial era are CO<sub>2</sub>, CH<sub>4</sub>, N<sub>2</sub>O and halocarbons. Systematic measurements of well-mixed GHGs at ambient air concentrations began at different times within the last six decades and expanded to a global monitoring network. The measurements of atmospheric CO<sub>2</sub> started at Mauna Loa, Hawaii, USA in 1958 and established atmospheric mole fraction of 315 ppm in 1958 (Keeling et al. 1976). Direct atmospheric CH<sub>4</sub> measurements of sufficient spatial coverage to calculate the global annual means began in 1978 (Dlugokencky et al. 1994), while that of atmospheric N<sub>2</sub>O started in late 1970s.

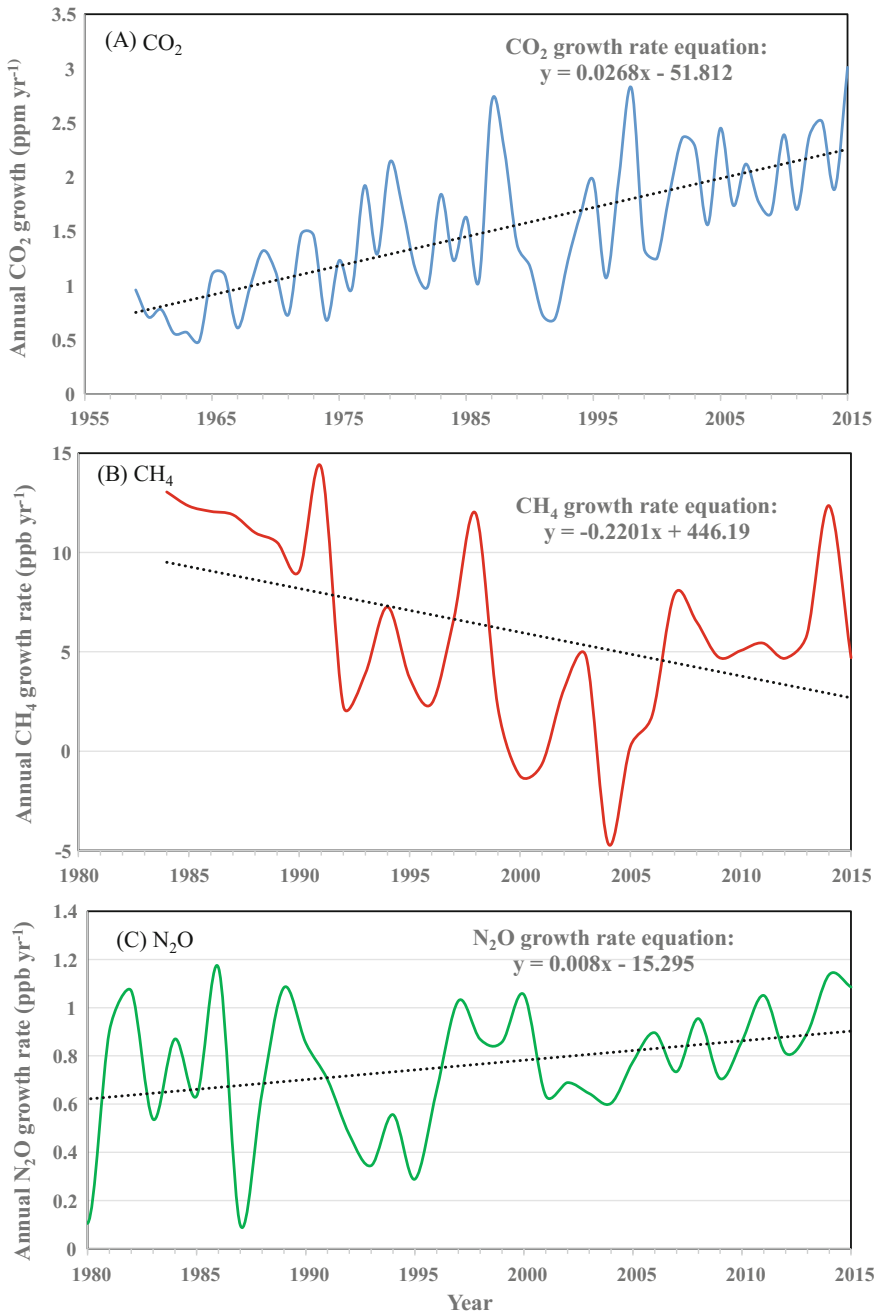
The global mean abundance of three major anthropogenic GHGs in 2015 were  $400.0 \pm 0.1$  ppm,  $1845 \pm 2$  ppb, and  $328.0 \pm 0.1$  ppb for CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O, respectively (WMO 2016). These values constitute 144, 256 and 121% abundances relative to pre-industrial (i.e., year 1750), respectively, and mean absolute increase of  $2.30 \text{ ppm yr}^{-1}$ ,  $11.0 \text{ ppb yr}^{-1}$ , and  $1.0 \text{ ppb yr}^{-1}$  for CO<sub>2</sub>, CH<sub>4</sub>, and N<sub>2</sub>O, respectively during the last 10 years (WMO 2016). The main contributors to the increase in atmospheric CO<sub>2</sub> are fossil fuel combustion and land use change. The average annual increase in globally averaged CO<sub>2</sub> during the instrumental record ranges from  $0.49 \text{ ppm yr}^{-1}$  to  $3.01 \text{ ppm yr}^{-1}$  (Fig. 2.2a). About 40% of CH<sub>4</sub> is emitted to the atmosphere by natural sources including wetlands, clathrates, wild ruminants, and termites, and 60% comes from anthropogenic sources such as domesticated ruminants, rice agriculture, fossil fuel exploitation, landfills and biomass burning. The average annual atmospheric CH<sub>4</sub> growth rate decreased from  $14.3 \text{ ppb yr}^{-1}$  in 1991 to near zero from 1999 to 2006. However, since 2007 atmospheric CH<sub>4</sub> has been increasing again (Fig. 2.2b), and its global annual mean increased by  $11 \text{ ppb yr}^{-1}$  between 2012 and 2015 (WMO 2016). The N<sub>2</sub>O is emitted from both natural (60%) and anthropogenic (40%) sources, which includes oceans, soils, biomass burning, fertilizer use in agriculture, and some industrial processes. Its increase in annual mean for the past 10 years is  $1.0 \text{ ppb yr}^{-1}$  (Fig. 2.2c). Other anthropogenic GHGs include chlorofluorocarbons (CFCs) and halogenated gases which are also ozone depleting compounds, sulfur hexafluoride (SF<sub>6</sub>) produced by chemical industry—mainly as an electrical insulator in power distribution equipment, hydrochlorofluorocarbons (HCFCs), and hydrofluorocarbons (HFCs) produced by human activities from various industrial sources. Their

current atmospheric concentrations are presented in Table 2.1. While CFCs and most halons are decreasing as a result of regulation of their use under Montreal Protocol, HCFCs and HFCs are increasing at a rapid rates, although they are still low in abundance (WMO 2016; NOAA/ESRL 2016).

Anthropogenic land cover and land use changes are important at local and regional level, but their influence at global level has become less important over the past century (Hansen et al. 2007). Anthropogenic land cover change exerts direct impact on Earth radiation budget through change in surface albedo, modification of surface roughness, and latent heat flux. Land use change, particularly deforestation also has significant effect on well-mixed GHGs emission. Changes in solar irradiance is a natural phenomenon caused by dark regions on the solar disk (Sunspots) which causes slight variations in radiation from the Sun at about 11-year cycles (Seinfeld 2011). The direct radiative forcing as a result of increase in total solar irradiance since 1750 is estimated to contribute an RF ranging from 0.0 to +0.10  $\text{W m}^{-2}$  (Myhre et al. 2013; Table 2.1) and the amplification of changes in total solar irradiance by climate system is estimated to cause 20-year lag in climate response (Eichler et al. 2009).

Atmospheric aerosols, both natural and anthropogenic, generally originate from emissions of particulate matter or formation of secondary particulate matter from atmospheric gaseous precursors. The main constituents of the atmospheric aerosols are  $\text{SO}_4^{-2}$ ,  $\text{NO}_3^{-}$ ,  $\text{NH}_4^{+}$ , sea salt, black carbon (BC), dust, and primary biological aerosol particles (Boucher et al. 2013). Some aerosols increase atmospheric reflectivity of incoming solar radiation and others, such as BC are strong absorbers of energy and also modify shortwave radiation (Heald et al. 2014). Aerosols can also influence clouds albedo by serving as cloud condensation nuclei or ice nuclei. Atmospheric aerosols are some of the most uncertain driver of global climate change, because they can scatter or absorb radiation, thereby cooling or warming the Earth and the atmosphere directly (Heald et al. 2014). The overall impact of present-day atmospheric aerosols is estimated to be cooling, thereby counterbalancing some of warming associated with GHGs (Table 2.1; Myhre et al. 2013).

Cloud may be composed of liquid water (possibly in a super-cooled form), ice, or both (i.e., mixed phase). Clouds cover about two thirds of the globe (Stubenrauch et al. 2013; Stengel et al. 2015). Satellite data estimates that global annual short-wave cloud radiative effect of approximately  $-50 \text{ W m}^{-2}$  and greenhouse effect contribution of  $+30 \text{ W m}^{-2}$  through long wave radiative effect (Loeb et al. 2009; Stephens and L'Ecuyer 2015), implying a net cooling effects of clouds. The net downward flux of radiation at the surface is sensitive to vertical and horizontal distribution of clouds, however. In addition to increasing albedo and causing cooling of the planet, clouds can also exert radiative effect at the surface and within the troposphere, thus influencing the hydrological cycle and circulation. Overall, the net effect of clouds to climate depends on its physical properties—such as level of occurrence, vertical extent, water path, nature of cloud condensation nuclei population, and effective cloud particle size (Boucher et al. 2013).



**Fig. 2.2** Instantaneous growth rates for globally averaged atmospheric CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O since the instrumental record for each of the major greenhouse gases. Data from NOAA/ESRL

In addition to a well-known anthropogenic greenhouse effect associated with emission of GHGs (e.g., CO<sub>2</sub>, CH<sub>4</sub>, N<sub>2</sub>O, and CFCs) humans also enhance greenhouse effect by emission of pollutants such as CO, volatile organic compounds (VOCs), nitrogen oxides (NO<sub>x</sub>), and SO<sub>2</sub>. Although these air pollutants are negligible in the atmosphere to cause any significant greenhouse effect, they have indirect greenhouse effect by altering the abundance of important GHGs such as CH<sub>4</sub> and ozone (O<sub>3</sub>) through their atmospheric chemical reactions and also acting as precursors of tropospheric O<sub>3</sub> and aerosols formation. They also impact atmospheric OH concentrations as well as CH<sub>4</sub> atmospheric lifetime (Hartmann et al. 2013). The main sources of atmospheric CO are direct emission from incomplete combustion of biomass and fossil fuels, as well as in situ production by oxidation of CH<sub>4</sub>. As a result, the atmospheric chemistry and climate change tend to be intrinsically linked. Humans also affect water budget of the planet by changing land surface, resulting into redistributing latent and sensible heat fluxes. Land use changes such as conversion of forests to cultivated land, changes in characteristics of vegetation, change in land surface color through burning, etc. changes reflectivity of the land (i.e., surface albedo), and also the rates of evapotranspiration and IR emissions.

### 2.4.1 *Signs of Changing Climate*

Natural forcings have contributed to climate change in past, such as glacial-interglacial cycles. However, the observed climate change in post-Industrial Era, and especially since 1950s has been attributed to other external changes, especially the change in composition of the atmosphere during the industrial period as a result of anthropogenic activities. Many aspects of the global climate are changing rapidly, and there is a wealth of observational evidence that climate is changing, such that warming of the climate system in recent decades is unambiguous. The IPCC in their recent assessment (AR5) concluded that "...warming of the climate is unequivocal, and since the 1950s many of the observed changes are unprecedented over decades to millennia. The atmosphere and ocean have warmed, the amount of snow and ice have diminished and sea level has risen" (p 40, Synthesis Report, IPCC 2014). The atmospheric CO<sub>2</sub> and CH<sub>4</sub> concentrations depart from Holocene and even Quaternary patterns from 1850 with more markedly changes from 1950, with the associated fall in  $\delta^{13}\text{C}$  as captured by tree rings and calcareous fossils (Waters et al. 2016).

Although the warming of the Earth's surface is the most cited evidence of the climate change, wide range of observations and lines of evidence for climate change exist. These includes: (i) atmospheric surface: air temperature, precipitation, air pressure, water vapor, wind speed; (ii) Atmospheric upper air: earth radiation budget, temperature, atmospheric water vapor, wind speed and direction, cloud properties; (iii) atmospheric composition: GHGs, ozone, other long-lived gases, aerosols and their precursors; (iv) ocean surface: temperature, salinity, sea level, sea



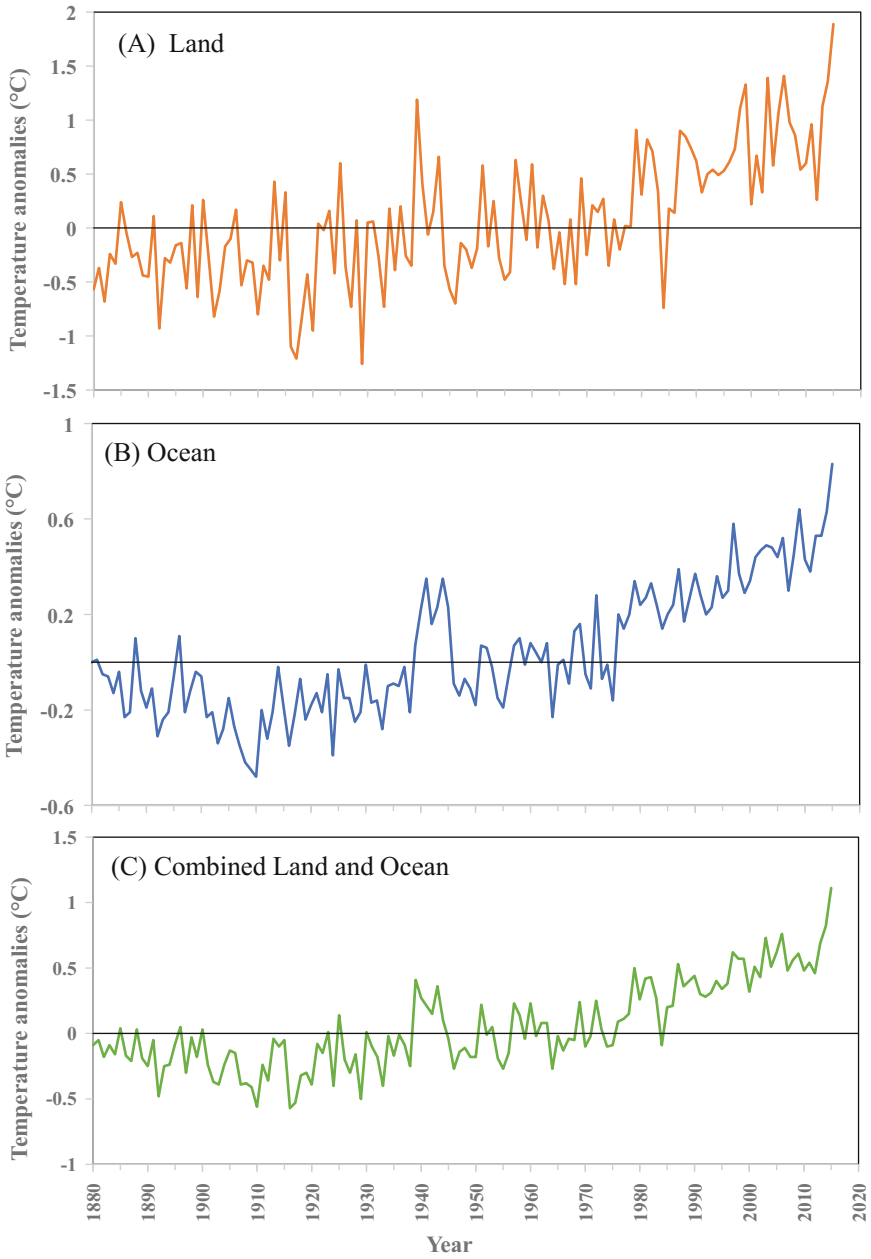
ice, ocean current, phytoplankton, CO<sub>2</sub> uptake, ocean acidity, nutrients, oxygen concentration; (v) ocean subsurface: temperature, salinity, subsurface water current, C composition; (vi) terrestrial: snow cover, albedo, permafrost, glaciers and ice caps, photosynthetically active radiation, water use efficiency (WUE), land cover, changes in hydrological cycle—soil moisture, river discharge, ground water (Hartmann et al. 2013; Arndt et al. 2015). Observational evidence of changing climate system has been obtained from multiple independent climate indicators, from the top of atmosphere to the depths of oceans. It includes the cooling of lower stratosphere, warming of lower troposphere, warming of the Earth and ocean surface, and increasing heat content of the ocean. Other changes include those in ocean temperatures, glaciers, snow cover, sea ice, sea level, and atmospheric water vapor (Arndt et al. 2015).

The global land temperatures at the surface, in the troposphere (i.e., the active weather layer extending about 8–16 km above the surface) and in the oceans have all increased in recent decades. The global land surface air temperatures has increased over the period of the instrumental record with warming rates approximately doubling since 1979 (Fig. 2.3a). Together with record-high GHG concentrations, the annual global surface temperature is currently the highest it has ever recorded for the period of 135 years of modern record keeping (Arndt et al. 2015). Several independently analyzed global and regional land surface air temperature (LSAT) data show only minor perturbations to global LSAT records since 1900, and revealed consistently increasing decadal LSAT anomaly trend (Table 2.2; Jones 2016). Furthermore, changes in surface atmospheric specific and relative humidity over the period are physically consistent with the reported global observed temperature trends (Peterson et al. 2011; Simmons et al. 2010).

The global average sea surface temperatures (SST) have increased since the beginning of the 20th century, as revealed by records obtained by different measurement methods (Fig. 2.3b; Kennedy et al. 2012). Although prominent spatiotemporal structures such as El Niño South Oscillation (ENSO) and decadal variability patterns in the Pacific Ocean exist, since 1950, SST has increased in all latitudes over each ocean. Different methods have been used to monitor SST over time, including moving ships, buckets, and satellite monitoring, and interpolation of the existing data by modeling. Analysis of these independently collected data show consistently increasing decadal SST anomaly trend (Table 2.2).

The global combined mean land surface and ocean surface temperature (GMST) has increased, with the last 50 years at almost double the rate of the last 100 years (Fig. 2.3c; Jones et al. 2012b; Morice et al. 2012). The GMST calculated by a linear trend has revealed a warming of about 0.85 °C (0.65–1.06 °C) from 1880 to 2012 and almost the whole globe has experienced surface warming with some decadal inter-annual variability (Hartmann et al. 2013). From 1980 each decade has been significantly warmer at the Earth's surface than the preceding decade (Table 2.2). Warming in the last century has occurred in two phases (Fig. 2.3c): (i) from 1910 to 1940s by about 0.35 °C, and (ii) more strongly from 1970s to present by about 0.50 °C. The past three decades have been warmer than all previous decades in the instrumental record. Also, World Meteorological Organization (WMO) identified





**Fig. 2.3** Global annual temperature anomalies from 1880 to 2015. Data source: <https://www.ncdc.noaa.gov/cag/time-series/global>

**Table 2.2** Trends of estimated land surface air temperatures (LSAT), sea surface air temperature (SSAT) and combined land and sea surface temperature (i.e., global mean surface temperature (GMST)).

Dataset	Reference	Temperature anomaly trend in °C decade <sup>-1</sup>				
		1880–2012	1901–2012	1901–1950	1951–2012	1979–2012
<i>Land surface air temperature (LSAT)</i>						
CRUTEM4	Jones et al. (2012b), Jones (2016)	0.086 ± 0.015	0.095 ± 0.020	0.097 ± 0.029	0.175 ± 0.037	0.254 ± 0.050
GHCN	Lawrimore et al. (2011)	0.094 ± 0.016	0.107 ± 0.020	0.100 ± 0.033	0.197 ± 0.031	0.273 ± 0.047
GISS	Hansen et al. (2010)	0.095 ± 0.015	0.099 ± 0.020	0.098 ± 0.032	0.188 ± 0.032	0.267 ± 0.054
Barkeley	Rohde et al. (2013)	0.094 ± 0.013	0.111 ± 0.034	0.111 ± 0.034	0.075 ± 0.029	0.254 ± 0.049
<i>Sea surface temperature (SST)</i>						
HadSST3	Kennedy et al. (2011), Atkinson et al. (2014)	0.054 ± 0.012	0.067 ± 0.013	0.177 ± 0.028	0.074 ± 0.027	0.124 ± 0.030
HadSST2	Rayner et al. (2006), Gil-Alana (2015)	0.069 ± 0.015	0.069 ± 0.012	0.084 ± 0.055	0.098 ± 0.017	0.121 ± 0.033
HadISST	Rayner et al. (2003)	0.042 ± 0.007	0.052 ± 0.007	0.067 ± 0.024	0.064 ± 0.015	0.072 ± 0.024
COBE-SST	Ishii et al. (2005)	–	0.058 ± 0.007	0.066 ± 0.032	0.071 ± 0.014	0.073 ± 0.020
ERSST	Smith et al. (2008)	0.054 ± 0.015	0.071 ± 0.011	0.097 ± 0.050	0.088 ± 0.017	0.105 ± 0.031
HadSST3	Kennedy et al. (2011)	0.054 ± 0.012	0.067 ± 0.013	0.117 ± 0.028	0.074 ± 0.027	0.124 ± 0.030
<i>Global mean surface temperature (GMST)</i>						
HadCRUT4	Morice et al. (2012), Jones (2016)	0.062 ± 0.012	0.075 ± 0.013	0.107 ± 0.027	0.106 ± 0.027	0.155 ± 0.033
NCDC	Vose et al. (2012), Jones (2016)	0.064 ± 0.015	0.081 ± 0.013	0.097 ± 0.010	0.118 ± 0.021	0.151 ± 0.037
GISS	Hansen et al. (2010)	0.065 ± 0.015	0.083 ± 0.013	0.090 ± 0.034	0.124 ± 0.020	0.161 ± 0.033

the year 2015 as the hottest on record, breaking all previous records by a margin of  $0.76 \pm 0.1$  °C above the 1961 to 1990 average, and for the first time, 2015 temperatures were 1 °C above the preindustrial era. In addition, years 2011–2015 were the warmest 5-year period on record (WMO 2015). Many extreme weather events, including heat waves which influences climate were also observed during this 5-year period.

The data show tropospheric warming in tropics, southern and northern hemispheres from 1958 to 2014 (Haimberger et al. 2012; Sherwood and Nishant 2015; Arndt et al. 2015) suggesting that atmospheric warming that has kept pace with global surface warming, while the stratosphere has been cooling during the same period (Santer et al. 2013; Sherwood and Nishant 2015; Arndt et al. 2015). The observed cooling in the stratosphere, while troposphere and global surface is warming is a fingerprint that the observed warming is due to increase in heat-trapping GHGs. In contrast, if the observed warming had been due to increases in solar output, Earth's atmosphere would have warmed throughout—global surface, troposphere and stratosphere (Santer et al. 2013, 2014). Other aspects climate, including changes in precipitation patterns (Min et al. 2011; Pall et al. 2011; Stott et al. 2016; Liu et al. 2016), increasing humidity (Santer et al. 2007; Willett et al. 2007; Mondal and Mujumdar 2015), change in pressure (Gillett and Stott 2009; Stott et al. 2016), and increase in heat content of the ocean (AchutaRao et al. 2006).

Natural drivers of climate cannot explain the observed warming, since over the last five decades, natural climate factors—such as solar forcing and volcanic eruptions—alone would have led to slight cooling (Gillett et al. 2012). The majority of the warming at the global scale over the past fifty years can only be explained by the anthropogenic influences, especially emissions from fossil fuels combustion (Santer et al. 2013; Stott et al. 2010). Consistent with scientific understanding of polar amplification of surface air temperature variations (Bekryaev et al. 2010), the largest increases in temperature are occurring close to the poles especially in the Arctic, and it is causing significant snow and ice cover to decrease in most areas, including Arctic sea, while atmospheric water vapor is increasing, since warmer atmosphere can hold more water. Global sea levels are also rising. Averaged over the recent decades, the sea levels are substantially different than they were half a century earlier.

Globally averaged surface air temperature has slowed its rate of increase since late 1990s (Fig. 2.3) even though each decade has been warmer than the previous. The slower recently observed rates of global surface warming from 2009 to 2012 which has been referred to as '*warming hiatus*' (Fig. 2.3) has been attributed to combination of factors, including cooling effects from natural radiative forcing (Santer et al. 2014), and energy redistribution within the ocean due to unforced variability (Palmer and McNeill 2014). It is estimated that more than 90% of increase in energy of the climate system between 1971 and 2015 has accumulated in oceans (Levitus et al. 2012; NOAA 2016), which goes towards warming the ocean (Abraham et al. 2013) and only about 1% of energy is stored in the atmosphere. Globally, the upper 0–75 m depth of the ocean water warmed by an average of  $0.11 \pm 0.2$  °C decade<sup>-1</sup> over the period 1971–2012 (Levitus et al. 2012). The

amount of heat accumulating in ocean plays vital role for diagnosing Earth's energy imbalance and sea level rise. Over the past four decades, the process of ocean heat uptake has resulted in marked increase in upper ocean heat content and ocean thermal expansion, thus contributing to sea level rise (Hanna et al. 2013). The ocean heat acts as buffer to climate change by slowing the rate of surface warming. The top 700 m depth ocean heat uptake from 1970 is estimated at  $19 \times 10^{22}$  J, which implies an average ocean water warming of 0.2 °C for the 43-year period (Abraham et al. 2013). The global sea level rise attributed to thermal expansion is 3 mm yr<sup>-1</sup> over the past 20 years (Abraham et al. 2013).

Cumulative CO<sub>2</sub> emissions from fossil fuels combustion, cement production, and land use change during the Industrial Era (1750–2015) is estimated at  $600 \pm 55$  Pg C, of which,  $260 \pm 5$  Pg C or ~43% remained airborne (Le Quéré et al. 2015, 2016), resulting into ~122 ppm increase or a relative abundance of ~144% compared to the 1750 atmospheric concentrations (WMO 2016). An estimated  $175 \pm 20$  Pg C or ~29% was taken up by the global oceans, while the remaining balance is believed to be retained in the terrestrial sinks (Le Quéré et al. 2015, 2016). Enhanced ocean uptake of CO<sub>2</sub> alters the marine CO<sub>3</sub><sup>2-</sup> system that controls sea water acidity. Oceans have absorbed 25–30% of the anthropogenic CO<sub>2</sub> emissions since the beginning of the industrial revolution. Although oceans act as C sinks, the CO<sub>2</sub> absorption process has a direct and measurable impact on ocean chemistry. Since late 18th century, the average pH of surface waters has decreased by 0.1 units, from 8.2 to 8.1, which is equivalent to 30% increase in H<sup>+</sup> concentration (Logan 2010). Changes in ocean chemistry affects marine life by (i) affecting calcification process of calcifying organism e.g., corals, (ii) decreasing pH which affects acid-base regulation and other physiological processes in the ocean, and (iii) increasing dissolved CO<sub>2</sub> which could affect the ability of primary producers to photosynthesize. Ocean surface has been more impacted with CO<sub>2</sub> uptake than the deep ocean. The observed change in ocean chemistry that indicates that the uptake of CO<sub>2</sub> has led to a reduction of the pH of surface seawater by 0.1 units, equivalent to a 30% increase in the concentration of H<sup>+</sup> is dominantly associated with surface and near surface waters (Raven et al. 2005). Similarly, there are indications that, parallel to increase in acidity and ocean water warming, O<sub>2</sub> concentration has decreased in coastal waters, open ocean thermoclines since 1960s, with likely expansion of tropical O<sub>2</sub>-minimum zones in recent decades.

In addition to observed increase of the land and sea surface temperature in the last 100 years (Fig. 2.3), over the last 30 years, satellites have made it possible to observe much broader spatial distribution of measurements, and indicate that upper ocean temperature has increased since at least 1950 (Willis et al. 2010; Lyman et al. 2010; Roemmich et al. 2015). Observations from satellites and also in situ measurements have suggested reductions in glaciers, Arctic sea ice and ice sheets (Holland and Kwok 2012; Matear et al. 2015). In addition to imbalances in radiation budget discussed in Chap. 1, satellite datasets based on measurements of electromagnetic radiations suggest small imbalance in heat content of the ocean (Hartmann et al. 2013). Atmospheric water vapor is also increasing in the lower

atmosphere, because warmer atmosphere can hold more water. Changes in average conditions have been accompanied by increasing trends in extremes of heat and heavy precipitation events and decreases in extreme cold (Alexander et al. 2006). It has been predicted that a doubling of the Earth's atmospheric CO<sub>2</sub> concentration from preindustrial concentration would warm Earth's surface by an average of between 1.5 and 4.5 °C (Kiehl 2007; IPCC 2013). Globally, GHGs emissions have been increasing as the growing demand for energy has more than offset the progress made in improved energy efficiency and deployment of new energy sources with lower GHG emissions. Natural drivers of climate cannot explain the current observed warming.

The ability to predict manifestations of changing climate carries considerable uncertainties suggesting that it is quite possible that the climate change impacts may be considerably worse in the near future than predicted. One of the examples is the reduction in Arctic perennial ice sheet, which has diminished at a rate of 13% per decade relative to previous mean rate from 1979 to 2012 (Stroeve et al. 2012b), which far exceeded model predictions (Stroeve et al. 2012a) and serve as indication that climate change impacts may occur rather sooner than expected. A number of other manifestations of the changing climate have been also observed, including rising sea level, drought, heat waves, more severe storms, increasing precipitation intensity, and associated disruption of terrestrial and aquatic ecosystems. In addition, increased atmospheric CO<sub>2</sub> concentration is diffusing into the ocean and acidifying surface waters and affecting marine ecosystems. Natural processes are currently removing less than half of anthropogenic CO<sub>2</sub> emissions from the atmosphere each year.

### ***2.4.2 Climate Change Metrics***

Efforts to mitigate anthropogenic climate change need to be able to assess the relative effectiveness of measures addressing the different forcing agents. Metrics are used in studies of climate change to simplify interpretation of the complex feedbacks and interactions that determine the ultimate effect of forcings. Various metrics of climate influence have been developed, each with its advantages and disadvantages. A climate change metric, generally is a variable or set of variables designed to parameterize a set of known or deduced influences on climate system that may result in climate change. The climate metrics is then used as a proxy to indicate the impact of forcing on the climate system resulting in change in energy balance of the Earth-Atmosphere system.

The potential uses of climate metrics include: (1) providing rapid evaluations of multiple potential approaches proposed to minimize the impact of anthropogenic activities on the climate system, (2) evaluation of relative contribution of two or more emissions from different anthropogenic activities to climate change, (3) evaluation of climate effects of competing technologies or energy usage and contributions of different emissions, (4) establishment of a basis for comparing

changes in climate effects in different countries or regions, (5) evaluation of proposed policies that encourage beneficial activities or discourage non-beneficial activities, (6) help industries or countries determine the best approaches and practices to meet specific commitments to reduce climatic impacts, and (7) quantification of the relative contributions of countries (Ravishankara et al. 2015). Some of required key features of a metric include: (i) scientifically sound, and also simple to use and easy to understand and communicate, (ii) applicable to scientific questions or policy issues of interest to the user, (iii) useful as a tool for communicating impact information among scientists, industry, and policy makers, (iv) transparent enough to convey the intended information by itself, and (v) simple, but creating confidence in the scientific integrity and trust of the metric. Among the various metrics, the more well-known are Global Warming Potential (GWP) and Global Temperature change Potential (GTP).

#### 2.4.2.1 Global Warming Potential

Each GHG differs in its atmospheric lifetime and radiative efficiency (i.e., how effectively a gas absorbs and re-radiates IR). The GWP is the most commonly used metric for comparing the RF of gases at different lifetimes, radiative characteristics and gas cycling. Using the GWP, GHG fluxes are converted to a common unit of ‘CO<sub>2</sub> equivalents’ and then compared directly to one another to determine whether ecosystem has a net warming or cooling effect on global climate. The GWP is defined as time-integrated RF due to pulse emission of given component, relative to a pulse emission of an equal mass of CO<sub>2</sub>. In other words, for the a given GHG, it is an index measuring RF following a pulse emission of a unit mass of a GHG in the present day atmosphere integrated over a chosen time horizon, relative to that of CO<sub>2</sub>. The GWP represents the combined effect of the differing times these gases remain in the atmosphere and their relative effectiveness in causing RF. Since GWP is a time integrated index, its value changes depending on the timescale of interest. The Kyoto protocol is based on GWPs from pulse emissions over 100-year time frame. The ecosystem ecologists and climatologists are interested in near term climate change, and therefore, most GHG calculations uses 100-year time frame. A direct interpretation of the existing definition is that GWP is an index of total energy added to the climate system by a component in question relative to that added by CO<sub>2</sub>. However, the GWP does not lead to equivalence with the temperature change or other climate variables due to differences in the atmospheric lifetimes of the gases (Daniel et al. 2012). Thus, the name GWP may be somewhat misleading, and relative cumulative forcing index may be more appropriate term (Myhre et al. 2013). The GWP is generally used as a default metric for transferring emissions of different gases to a common scale, often called ‘CO<sub>2</sub> equivalent emissions’. The GWP of a component  $i$  is expressed as [Eq. 2.2]:

$$GWP_i = \frac{\int_0^{TH} RF_i(t)dt}{\int_0^{TH} RF_r(t)dt} = \frac{\int_0^{TH} a_i \cdot [C_i(t)]dt}{\int_0^{TH} a_r \cdot [C_r(t)]dt} = \frac{AGWP_i}{AGWP_{CO_2}} \tag{2.2}$$

where, *TH* is the time horizon, (i.e., 20, 100, or 500 year horizon), *RF<sub>i</sub>* is the global mean RF of component *i*, *a<sub>i</sub>* is the RF per unit mass increase in atmospheric abundance of component *i* (i.e., radiative efficiency) [*C<sub>i</sub>(t)*] is the time dependent abundance of *i*, and corresponding quantities for the reference gas (*r*) in the

**Table 2.3** Strengths and limitations of global warming potential (GWP) and global temperature change potential (GTP)

Global warming Potential (GWP)	Global temperature change potential (GTP)
<b>STRENGTHS</b>	
<ul style="list-style-type: none"> <li>• The concept is easy to understand</li> <li>• The concept is easy to calculate</li> <li>• It is successful at transforming various GHGs to a common unit (CO2 equivalent)</li> <li>• It performs time integration of radiative forcing (RF) to project climate change to some future time</li> <li>• It can be modified to include equivalent forcing using efficacies</li> <li>• It is widely used in existing policy</li> </ul>	<ul style="list-style-type: none"> <li>• The concept is relatively simple and transparent</li> <li>• The concept requires few input variables</li> <li>• It allows calculation of time-dependent change in temperature which GWP does not allow</li> </ul>
<b>LIMITATIONS</b>	
<ul style="list-style-type: none"> <li>• It only considers effects for which RF is calculated</li> <li>• It does not evaluate the temperature change or time evolution of temperature change</li> <li>• Time integration of RF is not clearly defined</li> <li>• Comparison of short-lived or non-homogenous forcing is difficult (same problem for all existing metrics)</li> <li>• It incorporates all the limitations inherent in RF except that of atmospheric lifetime that is fully accounted for</li> <li>• Characterization of the impact of gas is not robust with respect to the climate impact</li> <li>• It is not a stable metric, GWP values have changed over time due to rapid improvements in understanding of the C cycle</li> <li>• It is difficult to what an appropriate time horizon should be, although the 100 -year has become the standard</li> <li>• It is less applicable in traditional configuration for fixed target policy analysis</li> </ul>	<ul style="list-style-type: none"> <li>• It may be limited to sustained emission applications</li> <li>• More studies of pulse emission effects are needed to assess its effectiveness</li> <li>• It depends on numerical value of climate sensitivity which is not well known</li> <li>• There is no clear choice on how to define equivalence</li> <li>• Similar to GWP and other emissions-based metrics, it is difficult to include non-emission-related effects including those occurring with formation of contrails</li> </ul>

denominator (Forster et al. 2007), *AGWP* is absolute global warming potential for gas *i* or  $CO_2$ . The *AGWP* is calculated by integrating the RF due to emission pulses over a chosen time horizon (i.e., 20 and 100-year). It can have advantages for certain applications because *AGWPs* are not dependent on comparisons with  $CO_2$ . Comparison with  $CO_2$  may not always be desired in some emissions (e.g., comparisons of  $NO_x$  emission effects from aviation relative to  $NO_x$  emissions from ground-based transportation systems).

*GWP* is a better measure of relative effects on climate than the comparison of RFs for different gases because *GWPs* differentiate between gases that would reside in the atmosphere for vastly different amount of times from few days to several centuries (Ravishankara et al. 2015). The *GWP* is approximately equal to the ratio of temperature response due to a sustained emission of species or the integrated temperature response for a pulse emission normalized to similar expression for  $CO_2$  (Azar and Johansson 2012; Ravishankara et al. 2015). Strengths and limitations of *GWP* are summarized in Table 2.3. Multiplying the mass of gas [in megagrams (Mg)] by the associated 100-year *GWP* derives the  $CO_2$  equivalent for a gas.

#### 2.4.2.2 Global Temperature Change Potential

An alternative to *GWP* that has received considerable attention recently is the Global Temperature change Potential (*GTP*), a metric that evaluates the cause-effect, and is defined as the change in global mean surface temperature at a selected point in time response to an emission pulse relative to that of  $CO_2$  (Neubauer and Megonigal 2015; Shine et al. 2005). Although *GTP* has received much attention as alternative to *GWP*, it has not yet been used for policy decisions. *GTP* gives the temperature changes as a function of time rather than that of integrated over certain time. *GTP* which was proposed by Shine et al. (2005), derive the relative temperature increase per unit mass of emission of GHG relative to that for an equivalent mass of emitted  $CO_2$  for a chosen integrated time horizon. *GTP* can be defined as either a pulse ( $GTP_p$ ) or a sustained emission ( $GTP_s$ ) can be used for calculating *GTP* (Shine et al. 2005). The *GTP* concept is based on assumption that the global mean surface temperature can be defined based on change in RF and heat capacity (*C*) as (Eqs. 2.3, 2.4):

$$\Delta T(t) = \frac{1}{C} \int_0^t \Delta F(t') \exp\left(\frac{t' - t}{\lambda C}\right) dt' \quad (2.3)$$

or:

$$C \frac{d\Delta T(t)}{dt} = \Delta F(t) - \frac{\Delta T(t)}{\lambda} \quad (2.4)$$



where,  $\Delta T$  is the change in temperature as a function of time,  $\Delta F$  is change in RF,  $C$  is the heat capacity of the mixed layer ocean,  $\lambda$  is the assumed climate sensitivity, the exponential is an impulse response function to forcing at some initial time  $t'$  and  $t$  is some time in the future. In this equation, it is assumed that ocean and land respond together and at the same rate, and therefore, they are together represented by single heat capacity. This assumption allows the climate system to have single time constant rather than slow time constant (ocean) and fast time constant (land), thus greatly simplifying the calculation.

For a known time-dependent increase or decrease in the concentration ( $S$ ) of a greenhouse gas, the concentration change over time is expressed as (Eq. 2.5):

$$\Delta X(t) = \alpha \Delta S \left[ 1 - \exp\left(-\frac{t}{\alpha}\right) \right] \quad (2.5)$$

where  $\alpha$  is the time constant for removal of gas  $x$ , and for the forcing ( $F$ ) given by  $A\Delta X(t)$ , absolute GTP for a sustained emission change ( $AGTP_s$ ) at a particular integration time for a forcing  $x$  is derived by [Eq. 2.6]:

$$AGTP_s^x = \frac{\alpha_x A_x}{C} \left\{ \tau \left[ 1 - \exp\left(-\frac{t}{\tau}\right) \right] - \frac{1}{\tau^{-1} - \alpha^{-1}} \left[ \exp\left(-\frac{t}{\alpha_x}\right) - \exp\left(-\frac{t}{\tau}\right) \right] \right\} \quad (2.6)$$

where  $\alpha$  is the time constant for removal of the gas  $x$ ,  $A$  is the RF for  $s$  1 kg change in concentration for gas  $x$ ,  $C$  is the heat capacity of the mixed layer ocean, and  $\tau$  is the time constant ( $\lambda C$ ) for the climate system. Similar to AGWPs, the AGTPs for  $CO_2$  are more complicated because of the complexity of its removal processes. The resulting GTP as a function of a gas or other forcing agent  $x$  is the ratio of AGTP for  $x$  divided by AGTP for  $CO_2$  [Eq. 2.7]:

$$GTP_s^x(t) = \frac{AGTP_s^x(t)}{AGTP_s^{CO_2}(t)} \quad (2.7)$$

and like GWP, GTP uses radiative forcing, but calculates the response of the surface temperature to the RF for the emitted gas. GTP is also expressed as a relative change for a gas compared to  $CO_2$ , [Eq. 2.8]

$$GTP_i(t) = \frac{AGTP_i(t)}{SGTP_{CO_2}(t)} = \frac{\Delta T_i(t)}{\Delta T_{CO_2}(t)} \quad (2.8)$$

Therefore, GTP accounts for the thermal inertia and response of the climate system to provide a relative measure of temperature responses for specific time horizon. The strengths and limitations of GTP and GWP as climate matrices are compared and contrasted in Table 2.3. The GTP for various gases can be used to

relatively weigh the effects of different emissions to obtain ‘CO<sub>2</sub> equivalents’ similar to those of GWPs. However, the concepts for constructing GWPs and GTPs are fundamentally different. GTPs can account for physically based processes such as climate sensitivity and ocean-atmosphere exchange of heat that cannot be done with GWPs. GTPs can also account for the slow response of the deep ocean and the resulting effects on the temperature response, which in effect prolongs the response from emissions beyond what is controlled by the lifetime of the gas. As a result, GTPs therefore, account for both the atmospheric response time scale of the gas under evaluation as well as response time of the climate system. GTP concept also has inherent uncertainties such as those associated with the climate sensitivity and ocean heat uptake (Table 2.3; Myhre et al. 2013; Olivie and Peters 2013). GTPs tend to have larger uncertainty range compared to the GWPs determined for the same gas (Ravishankara et al. 2015). GWPs are an integrated measure of the system, and recent analysis has shown that GWPs are useful measure of the energy entering the climate system (Olivie and Peters 2013), and the derived GWP for a gas depends only on the integral of the RF. While the GTPs saves as an instantaneous measure, the pathway of the forcing following emission of the gas is important (Ravishankara et al. 2015).

## 2.5 Conclusions

The global climate is changing, and the change is apparent across a wide range of observations, including global temperature increase and sea level rise. The global warming of the past century is primarily attributed to human activities during the Industrial Era, and the global climate is projected to continue to change over this century and beyond. The magnitude of observed climate change beyond the next few decades depends primarily on the amount of GHGs emitted globally and the sensitivity of Earth’s climate to those emissions. The global abundance of CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O in the atmosphere at the beginning of Industrial Era in 1750 were  $278 \pm 2$  ppm,  $722 \pm 25$  ppb, and  $270 \pm 7$  ppb for CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O, respectively, and increased to  $400.0 \pm 0.1$ ,  $1.845 \pm 0.002$ , and  $0.328 \pm 0.0001$  ppm in 2015 for CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O, respectively. Mean annual absolute increase during the last decade is estimated at  $2.08$  ppm yr<sup>-1</sup>,  $6.0$  ppb yr<sup>-1</sup>, and  $0.89$  ppb yr<sup>-1</sup> for CO<sub>2</sub>, CH<sub>4</sub>, N<sub>2</sub>O, respectively. The globally averaged combined land and ocean surface temperature revealed a warming of  $0.65$ – $1.06$  °C for the period 1880 to 2012, and the global average surface temperature in 2015 reached a symbolic and significant milestone of  $1$  °C above the pre-industrial era. In addition to global temperature, a wide range of climate variables have been monitored, covering the atmosphere, the terrestrial, the ocean and even the paleoclimate records, all pointing to the fact that the global climate is changing, with significant global warming, and the human activities have contributed significantly to these changes. The most

notable anthropogenic activity is the combustion of fossil fuels, which has released an estimated  $410 \pm 20$  Pg C from 1750 to 2015 from geologic reserves, resulting into an atmospheric CO<sub>2</sub> growth of  $260 \pm 5$  Pg C for the same period, together with significant changes in ocean chemistry and heat content. The second significant contributor is the land use conversion and deforestation (Refer Chap. 6). From 1750 to 2015, land use and land conversion and deforestation have released an estimated  $190 \pm 65$  Pg C. The decadal land use emissions during 2005 to 2014 is estimated at  $1.0 \pm 0.5$  Pg yr<sup>-1</sup>, while the terrestrial C sink is estimated at  $3.0 \pm 0.8$  Pg C yr<sup>-1</sup>. The decadal CO<sub>2</sub> emissions averaged over the last decade (2006–2015) revealed that 91% of total CO<sub>2</sub> emissions was caused by fossil fuels combustion and industry, whereas land use and land use change contributed 9% of total CO<sub>2</sub> emissions during the last decade. During the last decade, CO<sub>2</sub> emissions partitioned among atmosphere (44%), ocean (26%) and terrestrial land sink (30%).

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