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Theory of Increasing Greenhouse Gases

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Abstract

Unlike elsewhere on the globe, temperature in the tropics has increased systematically. From observed tropical temperature, numerical simulations have reproduced the observed evolution of atmospheric CO_2, including its annual cycle. Much the same has followed empirically from the observed covariance between tropical temperature and *net* emission of CO_2 - the component of emission that actually changes CO_2. Intensified over tropical land, the robust coherence between those observed features establishes that the changes of tropical temperature do not follow from changes of CO_2, but rather produce them.

Here, we investigate the physical mechanisms through which observed warming in the tropics can produce the observed evolution of CO_2. The conservation law governing atmospheric CO_2, supported by the observed temperature dependence of surface fluxes and observed temperature in the tropics, is used to calculate the time-varying response of CO_2 net emission.

Thermally-induced emission is found to account for the preponderance of CO_2 net emission and, thereby, for the vast majority of anomalous CO_2. Magnified over tropical land, it represents interannual intensifications of net emission, notably during episodes of El Niño. Represented equally well is the long-term intensification of net emission during the last half century. That intensification alone accounts for about half of the observed increase of atmospheric CO_2. Anomalous CO_2 introduced by the calculated net emission closely tracks the observed evolution of atmospheric CO_2.

Having similar dependence on temperature is methane. Like net emission of CO_2, net emission of methane is intensified by tropical warming. The simultaneous intensification of CO_2 and CH_4 emission provides a unified understanding of their joint increase, one that follows naturally from thermally-induced emission.

Keywords: Carbon cycle; greenhouse gases; thermally induced CO_2; anthropogenic CO_2; absorption time; mixing ratio; CH_4 emission; tropical temperature change.

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1. Introduction

Preceding studies performed an in-depth analysis on nuclear-perturbed carbon 14, a tracer of carbon dioxide, through which it determined the effective absorption rate of atmospheric CO_2 (Harde and Salby 2021; Salby and Harde 2021a, hereafter SH1). In concert with the conserva-

[†1] Murry Salby unexpectedly passed away, when the manuscript to this contribution was almost ready for submission. As the third and culminating paper of a trifecta, which comprehensively treats the increase of greenhouse gases in the atmosphere, it was a deep commitment to complete this publication, in memoriam of Murry Salby.

tion law governing atmospheric CO_2, the effective absorption rate, in turn, determines the instantaneous equilibrium level of "anthropogenic CO_2" - the perturbation of CO_2 introduced by anthropogenic emission. Representing an upper bound on anthropogenic CO_2, its equilibrium level is much too small for anthropogenic emission to be responsible for the observed increase of atmospheric CO_2.

A companion study then found that much of the observed evolution of *net* CO_2 emission, the component of emission that changes CO_2, is represented by thermally-induced emission: the perturbation of natural emission that accompanies changes of temperature - notably in the tropics, where temperature has increased systematically (Salby and Harde 2021b, hereafter SH2). Represented by thermally-induced emission are sporadic intensifications of net emission, like those observed during episodes of El Niño. Represented equally well is the long-term intensification of net emission, which accelerated CO_2 growth during the last half century. Jointly, these components of net emission yield a thermally-induced perturbation of CO_2 that tracks the observed evolution of CO_2.

Both studies were inspired by numerical simulations which, on the basis of observed temperature, were able to reproduce the observed evolution of atmospheric CO_2, including its annual variation and stepwise increase (see: Harde 2017; Harde 2019; Harde and Salby 2021). Here, we explore the underlying physical mechanisms through which observed changes of temperature can produce the observed evolution of CO_2. Following an overview of the global equilibrium that controls atmospheric CO_2, Section 3 develops how the thermally-induced response to warming results through the conservation law that governs atmospheric CO_2. Section 4 applies the conservation law to calculate the thermally-induced response of net emission to observed warming, along with the resulting change of CO_2. The calculated evolution of CO_2 is then compared against its observed evolution. Finally, Section 5 briefly discusses the CH_4 increase with temperature analogous to CO_2.

2. Global Equilibrium of CO_2

Carbon dioxide in the atmosphere is controlled by a competition between its introduction through global-mean emission, E, and its removal through global-mean absorption, A. This physical constraint is expressed by the conservation law governing atmospheric CO_2:

$$\frac{dr_A}{dt} = E - A = E_{net}\,, \tag{1}$$

where r_A is the global-mean mixing ratio of atmospheric CO_2 and E_{net} is its global net emission into the atmosphere - the resultant of all contributions. As developed elsewhere (SH2), E_{net} is a small residual of the two competing influences, of order only a few % of E and A. Consequently, CO_2 remains in a state of quasi-equilibrium globally, wherein its emission and absorption approximately cancel one another. The conservation law then reduces to

$$E \cong A = \alpha_{eff} \cdot r_A\,, \tag{2.1}$$

where α_{eff} is the effective rate of CO_2 absorption. In (2.1), absorption of CO_2 is proportional to the instantaneous abundance of CO_2. Ubiquitous among physical systems, this form of damping is an observed feature of CO_2. It is manifest in the exponential decline of its nuclear-perturbed isotope $^{14}CO_2$, a tracer of overall CO_2 (Harde and Salby 2021; SH1).

Although global emission and absorption remain nearly in balance, their equilibrium can nonetheless drift through changes of emission. By changing atmospheric CO_2, changes of emission induce parallel changes of absorption (2.1), which maintain global CO_2 in quasi equilibrium. According to (1), however, the induced changes of r_A follow, not from E directly, but from its small residual, E_{net}, which determines anomalous CO_2.

The quasi-equilibrium that controls CO_2 globally does not apply locally. Local net emission

($e-a$) is equal to CO_2 flux that is observed locally. Plotted in Fig. 1 as a function of latitude, local net emission is positive in the tropics, over ocean as well as over land (Wanninkhof 1992; Palmer et al. 2019). To maintain global equilibrium, net emission has opposite sign in the extratropics, becoming negative poleward of $\Phi = 20°$:

$$e - a > 0 \quad \text{for} \quad |\phi| < 20°; \qquad e - a < 0 \quad \text{for} \quad |\phi| > 20° . \tag{2.2}$$

Figure 1: Local CO_2 flux evaluated from CO_2 partial pressure of ocean surface and atmosphere, averaged from estimates of Wanninkhof and McGillis (1999) and Nightingale et al. (2000).

Therefore, net emission integrated over the tropics, $E_T = \int (e - a)\, dS_x$, comprises global emission of CO_2 in (1):

$$E = E_T . \tag{2.3}$$

Likewise, net emission integrated over the extratropics, $A_x = -\int (e - a)\, dS_x$, comprises global absorption of CO_2:

$$A = A_X . \tag{2.4}$$

Global quasi-equilibrium is thus expressed by

$$E_T \cong A_X = \alpha_{eff} \cdot r_A . \tag{3}$$

Further, because CO_2 is well mixed over the atmosphere, on time scales longer than a couple of months, it is largely invariant with latitude.

If subjected to a positive perturbation of emission, $\delta E_T > 0$, the Earth-Atmosphere System (EASy) will adjust to re-establish equilibrium. It does so by increasing r_A until the ensuing perturbation of absorption, $\alpha_{eff} \cdot \delta r_A$, just balances the perturbation of emission, δE_T.

Relationship to Surface Properties

In the tropics, total emission of CO_2 has contributions from Ocean and Land. Those contributions follow from the respective fluxes, E_O and E_L, integrated over surface area:

$$\begin{aligned} E_T &= E_O + E_L \\ &= S\{S_O F_O + S_L F_L\}, \end{aligned} \tag{4}$$

where F_O and F_L denote areal-mean surface fluxes over ocean and land, S_O and S_L the fractional contributions of the tropics occupied by ocean and land, and S the total tropical area.

Over ocean, CO_2 flux is determined by the Atmosphere-Ocean contrast of CO_2:

$$F_O = G(k,s,Sc,v) \cdot \Delta r \,, \tag{5.1}$$

where

$$\Delta r = r_O - r_A \,, \tag{5.2}$$

r_O is the mixing ratio of ocean CO_2 (viz of air at equilibrium with ocean surface), G is an empirical function of gas transfer velocity k, solubility s, Schmidt number Sc, and wind speed v. Now, r_O in ppmv is equivalent to the partial pressure of CO_2 in μatm at the surface. Therefore,

$$\Delta r(\text{ppmv}) \Leftrightarrow \Delta p_{CO2}(\mu\text{atm}) \,, \tag{5.3}$$

Δr(ppmv) representing the atmosphere-ocean contrast of CO_2 partial pressure. In (5.1), the temperature dependence of solubility and Schmidt number offset one another (Wanninkhof 2014; Etcheto and Merlivat 1988). Consequently, ocean CO_2 flux (5.1) depends primarily upon wind speed and the atmosphere-ocean contrast of CO_2:

$$F_O \cong B(v) \cdot \Delta r \,. \tag{6.1}$$

Over land, CO_2 flux follows from soil respiration

$$F_L \cong R(T, \text{m}) \,, \tag{6.2}$$

which depends upon soil temperature T and moisture m. Soil respiration derives from microbial activity and upward diffusion of the produced CO_2 (cf. Maier et al. 2020). It is noteworthy that the partial pressure of CO_2 in soil vastly exceeds its partial pressure in the atmosphere. At 500 - 20,000 ppmv,

$$r_L >> r_A \,.$$

For a uniform perturbation of surface conditions, the enormous partial pressures of CO_2 found just a few tens of cm beneath the surface makes emission from land determinative in re-establishing equilibrium between the atmosphere and the Earth's surface.

Although R is influenced by soil moisture, it is controlled chiefly by surface temperature (Wood et al. 2013; Zhou et al. 2013), which is determined by air temperature:

$$F_L \cong R(T) \,. \tag{7}$$

The temperature sensitivity of soil respiration,

$$Q_L = \frac{1}{R}\frac{dR}{dT} \,, \tag{8}$$

is observed in the range 10%/K - 25%/K at temperate and polar latitudes (Reich and Schlesinger 1992; Lloyd and Taylor 1994). In tropical forest, however, copious precipitation and sunlight magnify active biomass, which supports soil respiration. There, the observed sensitivity to temperature is greatest (ibid; Brechet et al. 2018), 30%/K and higher (Wood et al. 2013; Nottingham et al. 2018).

3. Perturbation of Tropical Temperature

Consider now a positive perturbation of tropical surface temperature:

$$\delta T_T > 0 \,,$$

which is imposed and then maintained. Through the temperature sensitivity of surface fluxes, $\delta T_T > 0$ introduces a perturbation in emission, δE_T. It increases E_T, which is then no longer balanced by $A_X = \alpha_{eff} \cdot r_A$ in (2.1). EASy is thus driven out of global equilibrium, with positive net global emission (1). Initially, the atmosphere is unperturbed: $\delta r_A = 0$, so $E_{net} = \delta E_T > 0$. Net emission, however, increases atmospheric CO_2. Accompanying the increase of r_A is an increase of CO_2 absorption, $\alpha_{eff} \cdot \delta r_A$. Atmospheric CO_2 continues to increase until the perturbation of

absorption again balances the perturbation of emission, re-establishing global equilibrium (2.1). According to (3), (5), and (6), the initial and final equilibria satisfy the balances:

$$S\{S_O B(\nu)\cdot\Delta r + S_L R\} = \alpha_{eff}\cdot r_A \tag{9.1}$$

$$S\{S_O B(\nu)\cdot(\Delta r + \delta\Delta r) + S_L(R+\delta R)\} = \alpha_{eff}\cdot(r_A + \delta r_A)\,. \tag{9.2}$$

Subtraction yields the balance governing the perturbation:

$$S\{S_O B(\nu)\cdot\delta\Delta r + S_L\delta R\} = \alpha_{eff}\cdot\delta r_A\,, \tag{9.3}$$

where the LHS represents the perturbation of emission with $\delta\Delta r$ as the perturbation of the atmosphere-ocean contrast Δr and δR as perturbation of the respiration. Dividing (9.3) by (9.1) obtains

$$\frac{S_O B(\nu)\cdot\delta\Delta r + S_L\delta R}{S_O B(\nu)\cdot\Delta r + S_L R} = \frac{\delta r_A}{r_A}\,. \tag{10}$$

The fractional perturbation of atmospheric CO_2 equals the fractional perturbation of emission, independently of effective absorption.

3.1 Aqua Tropics: $S_L = 0$

In the absence of land, (10) reduces to

$$\frac{\delta r_A}{r_A} = \frac{\delta\Delta r}{\Delta r}\,, \tag{11}$$

or with (5.2)

$$\frac{\delta r_A}{r_A} = \frac{\delta r_O - \delta r_A}{r_O - r_A}\,. \tag{12}$$

Upon rearrangement, (12) yields

$$\frac{\delta r_A}{r_A} = \frac{\delta r_O}{r_O}\,. \tag{13}$$

The fractional increase of atmospheric CO_2 is identical to the fractional increase of ocean CO_2. The latter, in turn, is imposed though the increase of surface temperature, $\delta T_T = \delta T_O$.

Ocean surface temperature is represented in the Sea Surface Temperature Anomaly (SSTA). Plotted in Fig. 2, the record of tropical SSTA extends back through the onset of CO_2 measurements at Mauna Loa, albeit with reduced coverage before the satellite era. At the beginning of the Mauna Loa record, ocean surface in the tropics ($|\phi| < 20°$) had a temperature anomaly of

$$T_O(1959) \cong -0.2\,\mathrm{K}\,, \tag{14.1}$$

corresponding to an actual surface temperature of about 300 K (26.9°C). During the subsequent era of CO_2 measurements, tropical ocean temperature increased systematically, attaining a net increase in closing years of

$$\delta T_O \cong +0.9\,\mathrm{K}\,. \tag{14.2}$$

To investigate its impact, δT_O is imposed impulsively in 1959, the start of the Mauna Loa record, and thereafter maintained.[2]

[2] The equilibrium properties in (9) are state variables. Therefore, the transition between the two states is path independent and, hence, independent of whether δT_O is introduced gradually or impulsively.

Figure 2: Anomalous sea surface temperature in the tropic (20°S – 20°N). Source HadSST4.0 (Kennedy et al. 2019).

Chemical equilibrium of seawater is governed by carbonate chemistry. Involving dissolved ions, it is determined chiefly by the following reactions (eg., Zeebe 2012):

$$CO_2(g) \Leftrightarrow CO_2(aq) \tag{15.1}$$

$$CO_2(aq) + H_2O \Leftrightarrow H^+ + HCO_3^- \tag{15.2}$$

$$HCO_3^- \Leftrightarrow H^+ + CO_3^{2-}, \tag{15.3}$$

wherein the concentrations [ ] of species define the conserved families: Dissolved Inorganic Carbon (DIC), equivalent to total carbon, and Total Alkalinity (TA) as:

$$DIC = [CO_2(aq)] + [HCO_3^-] + [CO_3^{2-}] \tag{15.4}$$

$$TA = [HCO_3^-] + 2[CO_3^{2-}] + [B(OH)_4^-] + [OH^-] - [H^+] \tag{15.5}$$

with $B(OH)_4^-$ as tetrahydroxyborate. The reactions (15) are temperature dependent. So too then is the chemical equilibrium they describe. In addition to temperature (14.1), chemical equilibrium of the ocean surface is determined by the values of total alkalinity (e.g., Jiang et al. 2014)

$$TA = 2350\ \mu\text{mol/kg} \tag{16.1}$$

and dissolved inorganic carbon (e.g., Wu 2019)

$$DIC = 2000\ \mu\text{mol/kg}, \tag{16.2}$$

properties to which the ocean partial pressure of CO_2 is sensitive.[3]

Through (15), T_O, TA, and DIC determine ocean CO_2. At the start of the Mauna Loa record, chemical equilibrium defined by the above values had ocean CO_2 of

$$r_O = r_O(1959) \cong 354\,\text{ppmv}\,. \tag{16.3}$$

In the atmosphere, CO_2 then was

$$r_A = r_A(1959) \cong 315\,\text{ppmv}\,. \tag{16.4}$$

[3] In open ocean, TA varies by +/- 40 μmol/kg. At ocean boundaries it varies even more, by as much as 100 μmol/kg. Such changes have a huge impact on the ocean partial pressure of CO_2: At TA=2300 μmol/kg, a decrease of 40 μmol/kg increases r_O by 175 ppmv (e.g., Robbins et al. 2010). A decrease of 100 μmol/kg increases r_O by 580 ppmv.

The initial atmosphere-ocean contrast of CO_2 was therefore

$$\Delta r(1959) \cong 39\,\text{ppmv}. \tag{16.5}$$

From its initial temperature (14.1), tropical ocean surface warmed during the Mauna Loa era, resulting in the temperature perturbation (14.2) in 2020. This increase of temperature perturbed the ocean surface, driving it to a new chemical equilibrium. The perturbed equilibrium at the new temperature, $T_O + \delta T_O$, has the perturbed CO_2

$$r_O + \delta r_O \cong 367\,\text{ppmv}. \tag{17}$$

According to (16.3) and (17), the temperature perturbation increased ocean CO_2 by

$$\delta r_O \cong 13\,\text{ppmv}. \tag{18.1}$$

The increase of atmospheric CO_2 that eventually results from this increase of ocean CO_2 follows from (13):

$$\delta r_A = (13/354)\cdot 315 \cong 12\,\text{ppmv}. \tag{18.2}$$

Initially, the atmosphere was unperturbed: δr_A (1959) = 0, as illustrated in Fig. 3. The initial perturbation to atmosphere-ocean contrast of CO_2 was therefore

$$\begin{aligned}\delta\Delta r(t=0) &= \delta r_O(1959) - \delta r_A(1959)\\ &= \delta r_O(1959) = 13\,\text{ppmv}\end{aligned}. \tag{19.1}$$

In response to the surface perturbation, however, the resulting disequilibrium with the atmosphere generates positive net emission. Thereby, atmospheric CO_2 develops a perturbation, δr_A (Blue), which steadily increases. It, in turn, increases the back pressure exerted on the ocean surface, which opposes the ocean flux (6) by reducing the atmosphere-ocean contrast of CO_2 (Red). When equilibrium is re-established, the negative feedback of δr_A has increased sufficiently to reduce the perturbation in CO_2 contrast to:

$$\begin{aligned}\delta\Delta r(t\to\infty) &= \delta r_O - \delta r_A(t\to\infty)\\ &\cong 13-12\,\text{ppmv} \cong 1\,\text{ppmv}\end{aligned}, \tag{19.2}$$

now with a slightly smaller contrast of Δr = 38 ppmv and a final atmospheric mixing ratio caused only by the oceans of r_A = 327 ppmv.

Figure 3: Evolution of anomalous CO_2 in ocean (Green) and in atmosphere (Blue), along with atmosphere-ocean contrast of CO_2 (Red).

The final perturbation of Δr in the new equilibrium (19.2) is less than 8% of its initial perturbation (19.1). Hence, the atmosphere-ocean contrast of CO_2 changes little, a feature of the observed record (Fig. 4).[4] By eliminating most of the perturbation in Δr, the increased back pressure exerted by the atmosphere $\delta r_A(t \to \infty)$ simultaneously eliminates most of the perturbation in ocean emission, δE_O ; (4) - (5).

Figure 4: CO_2 mixing ratio observed at Hawaii in the atmosphere (Red) and ocean (Blue). Source: USGCRP (2017), Fig. 13.4.

The foregoing transition between equilibrium states establishes the sensitivity to temperature of ocean emission:

$$Q_O = \frac{1}{E_O}\frac{\delta E_O}{\delta T} = \frac{\delta \Delta r / \Delta r}{\delta T} \quad , \qquad (20)$$
$$= \frac{1/39}{0.9}\mathrm{K}^{-1} = 0.029\,\mathrm{K}^{-1}$$

About 3%/K, the sensitivity to temperature of ocean emission is much smaller than the sensitivity implied by the initial CO_2 contrast (13 ppmv): $Q_O \cong (13/39)/0.9 \cong 37\%/\mathrm{K}$.

3.2 Terrestrial Tropics: $S_O = 0$

The sharp reduction of perturbation emission by ocean (19) follows from its dependence on the resulting perturbation of atmospheric CO_2. Increased back pressure exerted by the atmosphere (δr_A) reduces the perturbation of the atmosphere-ocean contrast, $\delta\Delta r$, eventually limiting the perturbation in ocean emission to a small fraction of its initial perturbation.

Unlike ocean emission, soil respiration (7) has no such limitation - because R is not determined by the contrast of CO_2 between the atmosphere and surface. Rather, $R = F_L$ is determined by the vertical gradient of CO_2 flux beneath the surface, which is imposed from below by upward diffusion of CO_2-rich air through porous media (Maier et al. 2020). And, because CO_2 partial pressure in soil (r_L) vastly exceeds its partial pressure in the atmosphere, increased r_A has little impact on the CO_2 flux gradient and, hence, on respiration R that determines emission from land.

In the absence of ocean, (10) reduces to

$$\frac{\delta r_A}{r_A} = \frac{\delta R}{R}. \qquad (21)$$

The thermally-induced perturbation of respiration follows from the temperature sensitivity of R (8):

[4] Different to the central tropics the ocean-atmosphere flux at Hawaii is already slightly negative (see also Fig.1).

$$Q_L = \frac{1}{R}\frac{\delta R}{\delta T_L} \cdot \tag{22}$$

In contrast to ocean emission, the accompanying perturbation of land emission (7) is undiminished by increasing atmospheric CO_2; relative to thousands of ppmv in soil, increases of a few tens of ppmv in the atmosphere are immaterial. The limiting perturbation of atmospheric CO_2, δr_A ($t \to \infty$), is therefore controlled autonomously by the initial perturbation of emission from land, which is invariant during the adjustment to the new (i.e., perturbed) equilibrium.

The fractional increase of atmospheric CO_2 follows from (21) and (22) as

$$\frac{\delta r_A}{r_A} = Q_L \cdot \delta T_L \cdot \tag{23.1}$$

For finite temperature perturbation, (22) becomes $dR/dT = Q_L \cdot R$. If $Q_L \cong$ const, soil respiration increases with δT_L exponentially:

$$R(\delta T_L) = R(0) \cdot e^{Q_L \cdot \delta T_L} . \tag{23.2}$$

At extra-tropical latitudes, such dependence is, in fact, observed (Lloyd and Taylor 1994; Savage and Davidson 2001). However, at tropical latitudes, the range of temperature is so narrow that deviations from linearity

$$R(\delta T_L) \cong R(0) \cdot \left(1 + Q_L \cdot \delta T_L\right) \tag{23.3}$$

are not observed (see Fig. 5). In tropical forest, observations of soil respiration reveal temperature sensitivity in the range 25-35%/K (Wood et al. 2013; Nottingham et al. 2018; Brechet et al. 2017). Representative is the temperature dependence in Fig. 5, which reflects the temperature sensitivity of land emission:

$$Q_L \cong 0.32\,\mathrm{K}^{-1} >> Q_O \tag{23.4}$$

Figure 5: Temperature dependence of soil respiration observed in tropical forest. After Wood et al. (2013).

Measurements of R in Fig. 5 were conducted in Puerto Rico, which lies at the fringe of the tropics. However, biomass, through which soil respiration operates, is concentrated in the deep tropics: the Congo, Amazon basin, and the maritime continent in the western Pacific (Fig. 6). Although measurements are unavailable, temperature sensitivity there may be substantially greater.

The observed perturbation of tropical temperature during the Mauna Loa era (14.2) is represented in the record of SSTA at latitudes $|\phi| < 20°$ (see Fig. 2). It describes the temperature of ocean surface in the tropics. Satellite measurements by the Microwave Sounding Unit (MSU) suite of

instruments sample temperature over both ocean and land. The MSU record of tropical ocean is strongly coherent with the surface record of anomalous SST, but slightly reduced because the MSU record represents atmospheric temperature a couple of km above the surface. Much the same holds for the systematic increase of MSU temperature over tropical landmass (a trend of +0.16 K/decade) vs. over tropical ocean (a trend of +0.12 K/decade). Scaling the perturbation of ocean temperature in (14.2) accordingly provides perturbation temperature that is representative of tropical land:

$$\delta T_L = 1.2\,\mathrm{K} \tag{24}$$

Figure 6: Global distribution of above-ground biomass (Mg/ha), derived from satellite observations. Highlighted is Puerto Rico (Red), the measurement site of soil respiration in the tropics (Fig. 5). Source: European Space Agency.

This temperature perturbation increases land emission (7), which in turn increases atmospheric CO_2. The fractional increase of r_A follows from (23):

$$\frac{\delta r_A}{r_A} \cong 0.39\,. \tag{25.1}$$

From initial CO_2 of r_A = 315 ppmv (16.4), atmospheric CO_2 thus experiences a thermally induced increase of

$$\delta r_A \cong 122\,\mathrm{ppmv}\,. \tag{25.2}$$

This corresponds to ending CO_2 in 2020 of

$$\delta r_A \cong 437\,\mathrm{ppmv}\,. \tag{25.3}$$

3.3 Aqua and Terrestrial Tropics

The presence of both ocean and land couples emission from ocean to emission from land. The perturbation of ocean emission depends upon the perturbation of atmospheric CO_2, which in turn depends upon the perturbation of land emission.

From (10), the fractional increase of atmospheric CO_2 may be expressed

$$\frac{\delta E_O + \delta E_L}{E_O + E_L} = \frac{\delta r_A}{r_A}\,. \tag{26}$$

Although differing in surface area, tropical ocean and land have comparable total emission. According to marine observations of Δr (Takahashi 1997; Feely et al. 2001) and satellite observations of CO_2 over tropical land (Palmer et al. 2019), emission from tropical ocean and land surface lie in the range

$$0.63E_O < E_L < 4.57E_O . \tag{27}$$

With (26), it follows (Appendix A) that the fractional increase of atmospheric CO_2 must lie in the range:

$$0.15 < \frac{\delta r_A}{r_A} < 0.34 \cdot \tag{28}$$

The thermally-induced increase of atmospheric CO_2, collectively from tropical ocean and land, must therefore lie in the range

$$47\,\text{ppmv} < \delta r_A < 107\,\text{ppmv} . \tag{29.1}$$

This range of fractional increase corresponds to ending CO_2 during 2020 in the range

$$362\,\text{ppmv} < r_A < 422\,\text{ppmv} . \tag{29.2}$$

Observed CO_2 during 2020 was ~ 414 ppmv.

4. Response to Observed Warming

Over much of the Earth, evidence of a systematic and sustained change of temperature is scarce (SH2). What warming has occurred during the Mauna Loa era occurred sporadically, incoherent in time. At mid-latitudes, the MSU record reveals no systematic change. At polar latitudes, where a trend is evident, MSU retrievals are vulnerable to ice, snow, and high terrain, features that are prevalent at polar latitudes (see, e.g., Maslanic 2007; Wang et al. 2017; Cui et al. 2020; Swanson 2003).

In contrast, temperature in the tropics has increased systematically (Fig. 2). Evident in the records of both MSU and SSTA, this systematic warming is free of the uncertainty that surrounds microwave retrievals at polar latitudes. According to the development in Sec. 3, the systematic warming of tropical ocean and land surface must be attended by intensified net emission of CO_2. In fact, it is (SH2). Intensified net emission, in turn, must induce a drift in global equilibrium, with an attendant increase of CO_2.

Investigating the thermally-induced perturbation of CO_2 requires a record of tropical temperature that extends back to 1959, when Mauna Loa measurements began. Over ocean, this requirement is met by the record of SSTA (Fig. 2), albeit with limited accuracy before the advent of global coverage by satellites. Over tropical land, no such record is available. There, we must also rely on SSTA.

Ocean temperature is less coherent with the observed record of net CO_2 emission than is land temperature (ibid). Nevertheless, it is strongly coherent with land temperature, which is largely imposed by surrounding ocean. The strong interdependence of land and ocean temperature is evidenced during the satellite era, when there is complete coverage of the tropics. The two then vary co-dependently, achieving a correlation of 0.94. Despite the limitation of coverage, SSTA is the only means by which tropical temperature can be extended back to the onset of Mauna Loa measurements. Mindful of the limitations, we employ the record of SSTA to explore the thermally-induced perturbation of CO_2, which can then be compared against the observed evolution of CO_2. From (10), the fractional perturbation of emission equals the fractional perturbation of atmospheric CO_2. Hence,

$$\delta E = E\frac{\delta r_A}{r_A} , \tag{30.1}$$

where the perturbation of emission has contributions from ocean and land:

$$\delta E = E_O \cdot Q_O \cdot \delta T_O + E_L \cdot Q_L \cdot \delta T_L , \tag{30.2}$$

Then

$$\frac{\delta E}{E} = \frac{E_O \cdot Q_O \cdot \delta T_O + E_L \cdot Q_L \cdot \delta T_L}{E_O + E_L} = \frac{E_O(Q_O \cdot \delta T_O + \gamma \cdot Q_L \cdot \delta T_L)}{E_O(1+\gamma)}, \tag{31.1}$$

where

$$\gamma = \frac{E_L}{E_O} \tag{31.2}$$

represents the relative emission from tropical land and ocean (27). The fractional perturbation of emission then becomes

$$\frac{\delta E}{E} = \frac{Q_O \cdot \delta T_O + \gamma \cdot Q_L \cdot \delta T_L}{1+\gamma}. \tag{32}$$

Owing to the strong interdependence of temperature over tropical land and ocean, temperature over land can be expressed in terms of temperature over ocean. Proceeding as in Sec. 3.2, (24) obtains

$$T_L \cong (0.16/0.12)T_O = 1.33T_O \ . \tag{33}$$

The fractional perturbation of emission (32) then reduces to

$$\frac{\delta E}{E} = \frac{Q_O + 1.33 \cdot \gamma \cdot Q_L}{1+\gamma}\delta T = \overline{Q} \cdot \delta T, \tag{34.1}$$

where

$$\overline{Q} = \frac{Q_O + 1.33 \cdot \gamma \cdot Q_L}{1+\gamma} \tag{34.2}$$

represents the effective temperature sensitivity of tropical surface, collectively from ocean and land, and T is understood hereafter to refer to T_O .

For finite temperature perturbation and integration of (34.1) obtains the perturbation of overall emission from the tropics in terms of anomalous temperature of tropical ocean:

$$E(\delta T) = E(0) \cdot e^{\overline{Q} \cdot \delta T}, \tag{35}$$

where $E(0)$ is CO_2 emission in the unperturbed equilibrium state and δT is the departure of temperature from that state.

Anomalous emission in (35) describes the temperature-dependent part. Total emission is then

$$E = E_0 + \hat{E}(\delta T), \tag{36}$$

with E_0 representing emission that is not temperature dependent and $\hat{E}(\delta T)$ given by (35). Time-varying net emission, $\hat{E}(\delta T(t))$, can be evaluated over the Mauna Loa era from the contemporaneous record of SSTA: $\delta T(t)$.

4.1 Constant Effective Absorption

We are interested in determining E_{net} , the component of emission that controls anomalous CO_2. The conservation law (1) defines net emission,

$$E_{net} = E - \alpha_{eff} \cdot r_A, \tag{37.1}$$

which determines anomalous CO_2 through

$$E_{net} = \frac{dr_A}{dt}. \tag{37.2}$$

For constant α_{eff}, differentiating (37.1) obtains the balance governing net emission of CO_2:

$$\frac{dE_{net}}{dt} + \alpha_{eff} E_{net} = \frac{dE}{dt} . \tag{38}$$

Equation (38) can be solved analytically. Incorporating an integrating factor yields its solution:

$$E_{net}(t) = \left[E_{net}(0) - \hat{E}(0)\right] e^{-\alpha t} + \hat{E}(t) - \alpha\, e^{-\alpha t} \int_0^t e^{\alpha t'}\, \hat{E}(t')\, dt' , \tag{39.1}$$

wherein initial conditions are embodied in the term in brackets, α represents α_{eff}, and

$$\hat{E}(t) = \hat{E}(0) \cdot e^{\overline{Q} \cdot \delta T(t)} . \tag{39.2}$$

Although restricted to α = const, an analytical solution of E_{net} affords insight that is not available from a numerical solution. For times long compared to the effective absorption time ($t >> \alpha^{-1}$), net emission reduces to the limiting form

$$E_{net}(t) \sim \hat{E}(t) - \alpha\, e^{-\alpha t} \int_0^t e^{\alpha t'}\, \hat{E}(t')\, dt' . \tag{40}$$

Therein, net emission of CO_2 follows from a competition between temperature-induced emission, embodied in the first term on the right-hand side, and absorption of the resulting CO_2, embodied in the second term. In the latter, absorption discriminates according to the time scale of emission, through the exponential kernel of the integral in (40). Fluctuations of $\hat{E}$ on time scales shorter than α^{-1} experience strong cancellation in the integral. Thereby, those components are suppressed in the absorption term, leaving them largely undamped. If, however, α also discriminates according to time scale, as will emerge in Sec. 4.2, then short time scales can nevertheless experience strong absorption.

Notice: If the perturbation of temperature is invariant, δT= const,

$$\begin{aligned} E_{net}(t) &= \hat{E}(0) e^{\overline{Q} \cdot \delta T} - \alpha\, \hat{E}(0)\, e^{\overline{Q} \cdot \delta T}\, e^{-\alpha t} \int_0^t e^{\alpha t'}\, dt' \\ &= \hat{E}(0) e^{\overline{Q} \cdot \delta T} - \alpha\, \hat{E}(0)\, e^{\overline{Q} \cdot \delta T} \cdot \frac{1 - e^{-\alpha t}}{\alpha} \quad , \\ &\sim 0 \qquad\qquad \text{for } t >> \alpha^{-1} \end{aligned} \tag{41}$$

Irrespective of how large the temperature perturbation is, net emission eventually vanishes - because anomalous CO_2 eventually increases enough to restore the system to equilibrium. The same holds for any constant perturbation of emission (e.g., anthropogenic emission, were it to remain constant). It follows that, under these conditions, CO_2 cannot accumulate in the atmosphere indefinitely. Net emission can remain nonzero and, hence, continue to change CO_2, only if emission continues to change.

Because the global system remains in a state of quasi-equilibrium, wherein emission is largely balanced by absorption (2.1), induced changes of CO_2 must track changes of emission that produce them. These circumstances apply to thermally-induced CO_2, including the component forced by the systematic increase of tropical temperature (Fig. 2).

4.2 Arbitrary Absorption

For $\alpha = \alpha_{eff}$ variable, as emerges from the record of nuclear-perturbed carbon 14 (Harde and Salby 2021; SH1), differentiating (37) leads to the more general balance that governs net emission:

$$\frac{dE_{net}}{dt} = \frac{dE}{dt} - \alpha\, E_{net} - \frac{d\alpha}{dt} r_A . \tag{42}$$

With (36) and (37.2), this can be rearranged into

$$E_{net}(t) = \frac{1}{\alpha}\left[\frac{d\hat{E}}{dt} - \frac{dE_{net}}{dt}\right] - \frac{d\ln\alpha}{dt}\int_0^t E_{net}(t')\,dt' \,. \tag{43}$$

Governing net emission under conditions of arbitrary absorption, the balance (43) is an integro-differential equation of the Volterra class (Goltser and Litsyn 2005).

Anomalous surface temperature (Fig. 2) and, hence, thermally-induced emission (35) involve two spectral components of disparate time scales: (i) interannual fluctuations, with periods of a couple of years, and (ii) the systematic increase of temperature, with periods much longer than a decade. Because their time scales are widely separated, those spectral components correspond to high- and low-pass filtered forms of δT. Through (35), they introduce similar components into thermally-induced emission:

$$\hat{E}(t) = \hat{E}_{LP}(t) + \hat{E}_{HP}(t) \,. \tag{44}$$

Anomalous CO_2 produced by these spectral components of emission experiences different effective absorption, which depends upon time scale.

As developed in SH1, effective absorption of CO_2 is modulated by its re-emission from the Earth's surface. By offsetting direct absorption, re-emission slows effective absorption of CO_2. For short time scales (e.g., induced by interannual fluctuations of δT), re-emission has little time to intensify before δT reverses. Direct absorption of CO_2 then experiences little offset, leaving α_{eff} fast. However, for long time scales (e.g., induced by the systematic increase of δT), re-emission has ample time to intensify. Direct absorption of CO_2 then experiences a large offset, which renders α_{eff} slow.

The dependence on time scale is manifest in the response of CO_2 to $\hat{E}_{LP}$ and $\hat{E}_{HP}$. An oscillatory perturbation of temperature

$$\delta T = \mathcal{T}\, e^{-i\omega t} \tag{45.1}$$

and thus of emission

$$\delta E = \mathcal{E}\, e^{-i\omega t} \tag{45.2}$$

is introduced in the 3-volume system of SH1, a system that reproduces the observed evolution of nuclear-perturbed carbon 14 in the same way as seasonal oscillations. Through Fourier transform, spectral components of the form (45) comprise the LP and HP components of δT and, thereby, the respective components of thermally-induced emission, $\hat{E}_{LP}$ and $\hat{E}_{HP}$. For specified ω, the perturbation $\delta\mathcal{E}$ introduces an oscillatory perturbation in atmospheric CO_2. The amplitude of anomalous CO_2 measures effective absorption of that spectral component.

Anomalous tropospheric CO_2 in the 3-volume system (viz. r_A) is coupled to anomalous CO_2 in the surface layer and in the stratosphere through exchanges of carbon. Nevertheless, r_A still obeys the conservation law (1), which may be expressed abstractly

$$\frac{dr_A}{dt} + \alpha_{eff}\, r_A = F \,, \tag{46}$$

where the forcing operator F includes perturbation emission of CO_2, as well as exchanges of CO_2 between the troposphere, surface layer and stratosphere. Letting

$$\begin{aligned} F &= \mathcal{F}\, e^{-i\omega t} \\ r_A &= \mathcal{R}\, e^{-i\omega t} \end{aligned} \tag{47.1}$$

transforms (46) into

$$-i\omega\mathcal{R} + \alpha_{eff}\,\mathcal{R} = \mathcal{F} \,. \tag{47.2}$$

Anomalous CO_2 induced by the perturbation of emission (45.2) thus oscillates with the complex amplitude

$$\mathcal{R} = \frac{\mathcal{F}}{-i\omega + \alpha_{eff}} . \tag{48}$$

The perturbation of tropospheric CO_2 has the limiting form

$$\mathcal{R} \sim \begin{cases} \dfrac{\mathcal{F}}{-i\omega} = \dfrac{\mathcal{F}}{\omega} e^{i\pi/2} & \omega/\alpha_{eff} \to \infty \\ \dfrac{\mathcal{F}}{\alpha_{eff}} & \omega/\alpha_{eff} \to 0 \end{cases} . \tag{49}$$

Short time scales ($\omega \to \infty$) are introduced by $\hat{E}_{HP}$ (44). Anomalous CO_2 then becomes small and oscillates *in quadrature* with thermally-induced emission. In this limit, the troposphere and surface layer are out of equilibrium: CO_2 in one never equals that in the other. Re-emission from the Earth's surface is therefore weak. Consequently, it provides little offset to direct absorption, leaving α_{eff} fast.

Long time scales ($\omega \to 0$) are introduced by $\hat{E}_{LP}$. Anomalous CO_2 then oscillates *in phase* with thermally-induced emission. In this limit, the troposphere and surface layer have time to attain quasi-equilibrium, with CO_2 in one nearly equal to that in the other. Re-emission has then intensified. It offsets much of direct absorption, rendering α_{eff} slow.

Rearrangement of (48) obtains the effective absorption rate in terms of the amplitude of the spectral component with frequency ω:

$$\alpha_{eff}^2 = \frac{|\mathcal{F}|^2}{|\mathcal{R}|^2} - \omega^2 . \tag{50}$$

Note: Since $\mathcal{R}$ scales directly with $\mathcal{F}$, α_{eff} is independent of the forcing amplitude. Without loss of generality, $\mathcal{F}$ may therefore be set equal to 1 ppmv/yr.

Plotted in Fig. 7 is anomalous CO_2 that is induced by the spectral component of F with a period $2\pi/\omega = 2000$ yrs (Blue), which is inherent to the long-term increase in Fig. 2. Anomalous r_A that it produces is seen to vary in phase with its forcing, $F_{Re} = Re\{\mathcal{F}e^{-i\omega \cdot t}\} \sim \cos(\omega \cdot t)$. Related to $|\mathcal{F}|$ it has an amplitude of 9.6 ppmv. According to (50), the LP component of anomalous CO_2 therefore experiences effective absorption

$$\alpha_{eff}^{LP} \cong 0.104\ yr^{-1} , \tag{51.1}$$

or an effective absorption time $\tau_{eff} = 9.6$ yr in full agreement with Harde and Salby 2021 and SH1.

Also plotted in Fig. 7 is anomalous CO_2 that is induced by the spectral component of F with a period 2.8 yrs (Red), equal to the mean period of interannual fluctuations in Fig. 2. It is inherent to the HP component of thermally-induced emission. Despite being forced by the same amplitude in the same system, anomalous CO_2 introduced by the HP component of emission achieves an amplitude of only 0.37 ppmv. According to (50), it experiences effective absorption

$$\alpha_{eff}^{HP} \cong 1.51\ yr^{-1} , \tag{51.2}$$

corresponding to an effective absorption time of less than a year.

Figure 7: Response of tropospheric CO_2 to harmonic surface emission with a period of 2000 yrs (Blue) and a period of 2.8 yrs (Red), equal to the mean period of interannual fluctuations of net emission in the observed record: see text.

It is noteworthy that absorption of the HP component of anomalous CO_2 can also be deduced empirically. Interannual fluctuations of CO_2 are observed to operate coherently with but lag interannual fluctuations of temperature (Humlum et al. 2013; Salby 2013; Harde 2019). Independent of the above analysis, the observed lag reveals effective absorption of $\alpha_{eff}^{HP} \cong 1.35\ yr^{-1}$.[5]

The governing balance (43) is now solved numerically for thermally-induced net emission, $E_{net}(\delta T)$, through 4th order Runge-Kutta treatment. With $\hat{E}(t)$ prescribed through SSTA (39.2), anomalous temperature (Fig. 2) is filtered to periods longer and shorter than two decades, thereby discriminating between the LP and HP components of δT. Anomalous CO_2 introduced by the respective components of thermally-induced emission (39.2) is subjected to effective absorption in (51), which is varied weakly according to observed α_{eff} (SH1). Representative of the effective temperature sensitivity of tropical surface (34) is $\overline{Q} = 0.30\ K^{-1}$ (Fig. 5).[6] The corresponding value of γ in (31.2) is 2.7. It falls squarely within the observed range of emission inferred from observations (27).

Plotted in Fig. 8 is thermally-induced net emission (Blue). Despite having several hundred observations, of order a hundred independent degrees of freedom, $E_{net}(T)$ tracks the observed record of E_{net} (Red). The correspondence holds for interannual fluctuations of net emission, notably during the El Niños of 1973, 1997, and 2016, as well as for its long-term intensification, wherein E_{net} and $E_{net}(\delta T)$ intensify from ~ 0.7 ppmv/yr to almost 2.5 ppmv/yr. Notice that, before the El Niño of 1973, observed net emission remains approximately level: E_{net} undergoes no systematic change until after that El Niño. Thermally-induced net emission, $E_{net}(\delta T)$, undergoes the same transformation; cf. Fig. 2. With a correlation of 0.78, the two records in Fig. 8 are strongly interdependent. Their correlation is, in fact, quite close to the correlation between observed E_{net} and thermally-induced emission that is deduced empirically (SH2).

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[5] Consistent with both analyses is the observed annual cycle of emission from tropical land in the Northern Hemisphere, where its annual cycle is strongest (Palmer 2019). It varies nearly in phase with the annual cycle of CO_2. According to (49), this HP component of CO_2 must therefore experience effective absorption shorter than a year.

[6] In light of the paucity of measurements in the deep tropics, where biomass is concentrated, this value may well underestimate the actual temperature sensitivity.

Figure 8: Net CO_2 emission observed (Red), low-pass filtered to periods longer than a year, along with its thermally-induced component (Blue), calculated from the observed record of tropical temperature.

Subject to the same physical processes is anthropogenic emission, E^{Ant}, which is plotted in Fig. 9. It, too, must obey the governing balance (43), with E^{Ant} in place of $\hat{E}$. There is, however, an important distinction. Unlike thermally-induced emission from the surface, anthropogenic emission operates directly in the atmosphere. For re-emission of CO_2 to amplify and slow effective absorption of CO_2, anthropogenic emission must enrich CO_2 in the surface layer relative to that in the atmosphere. As detailed in SH1, the continuous re-supply of anomalous CO_2 in the atmosphere maintains the troposphere and surface layer out of equilibrium. Thereby, it limits re-emission and its offset of direct absorption, leaving α_{eff} fast.

Figure 9: Global anthropogenic emission of CO_2. Sourced from CDIAC (2017) and Le Quéré et al. (2021).

The 3-volume system of Fig. 7 is now perturbed by atmospheric emission with a period of 2000 yrs, characteristic of the long-term increase of anthropogenic emission (Fig. 9). Anomalous CO_2 (not shown) evolves similarly to the LP spectral component in Fig. 7, which results from surface emission. However, forced by emission in the atmosphere, r_A now achieves an amplitude of only 1.105 ppmv. From (50), anomalous CO_2 therefore experiences effective absorption of

$$\alpha_{eff}^{Ant} \cong 0.91\ yr^{-1}, \tag{51.3}$$

corresponding to an effective absorption time of about a year. Much the same effective absorption is found for monotonically-increasing anthropogenic emission (SH1).

Plotted in Fig. 10 is net emission (43) that follows from anthropogenic emission, E_{net}^{Ant} (Magenta, Lower Graph). Owing to rapid absorption, anthropogenic CO_2 is removed almost as fast as it is introduced. Consequently, E_{net}^{Ant} is only a few tenths of a ppmv/yr, an order of magnitude weaker than net emission from the Earth's surface that is thermally-induced, $E_{net}(\delta T)$; cf Fig. 8.

Figure 10: Net CO_2 emission observed (Red), along with the total calculated net emission, comprised of thermally-induced and anthropogenic components (Purple). Plotted separately is the anthropogenic component alone (Magenta, Lower Graph).

Superimposed in Fig. 10 is total net emission: $E_{net} = E_{net}(\delta T) + E_{net}^{Ant}$ (Purple). It tracks the observed record of net emission (Red). Total net emission follows almost entirely from the thermally-induced component (Fig. 8). Relative to mean net emission observed during the Mauna Loa era, the thermally-induced component represents 95.9%. The anthropogenic component represents 4.1%. Almost the same is expected from the natural to anthropogenic emission ratio.

Figure 11: Observed increase of CO_2 (Green), compared against the calculated increase (Purple). Note: Calculated CO_2 lies mostly beneath observed CO_2, but emerges near the end of the record. Plotted separately is the contribution from thermally-induced emission (Blue) and the contribution from anthropogenic emission (Magenta).

In Fig. 11 is displayed the observed increase of r_A (Green). Superimposed is anomalous CO_2 (37.2) that is forced by the total emission in Fig. 10 (Purple - almost completely covered by the Green graph). It closely tracks the observed evolution of anomalous CO_2. Tracking observed CO_2 almost as closely is the thermally-induced component of anomalous CO_2 (Blue). Also superimposed in Fig. 11 is the anthropogenic component of anomalous CO_2 (Magenta). Comparatively small, it tracks the instantaneous equilibrium level of anthropogenic CO_2 (Sec. 1) (see also Harde 2019, Figs 8 and 10; and SH1, Fig 14).

In this context we note that from estimates of the anthropogenic emission rate E^{Ant} and the observed increase $\Delta r_A/\Delta t$ of the atmospheric mixing ratio within one year, it is widely inferred that about 44% of these emissions (or an equivalent mass) as so-called Airborne Fraction AF remains in the atmosphere, while the rest is absorbed by extraneous reservoirs (Sixth Assessment Report (AR6), 2021, Chap. 5, Fig 5.7). Therefore, nature is assumed to be a net sink, and not natural but almost exclusively anthropogenic sources are made responsible for the increasing atmospheric CO_2.

However, this interpretation neglects that changes in the mixing ratio, which are initiated in a linearly responding system, cannot be greater than the perturbation, and these changes are responding with some delay to the emissions, determined by the effective absorption time τ_{eff}. So, comparing the growth rate $\Delta r_A/\Delta t$ of the mixing ratio with an additional emission ΔE, native or anthropogenic, it is obvious that $\Delta r_A/\Delta t < \Delta E$, until new equilibrium or quasi-equilibrium can establish with $\Delta r_A^{eq} = \Delta E \cdot \tau_{eff}$. But by far this does not mean that nature would be a net sink and could not contribute on its own to an increasing mixing ratio (see Appendix B).

For a more detailed discussion of the different paradigms of native or anthropogenic contributions to the CO_2 mixing ratio in the atmosphere, see also Connolly et al. 2020.

5. Methane Emission

Like carbon dioxide emission, CH_4 emission increases with temperature (Fig. 12). Methane has likewise been systematically increasing, albeit in an observed record that is much shorter than the Mauna Loa record of CO_2. It too is emitted by biomass (Fig. 6), chiefly through anaerobic processes that operate in well-irrigated regions like wetlands. Owing to its production in aqueous environments, however, methane emission is also dependent upon precipitation, which is plotted in Fig. 13a. It mirrors the observed distribution of methane, presented in Fig. 13b. Accordingly, those influences magnify CH_4 emission from tropical land, where biomass and precipitation are abundant (Pangala et al. 2017; Lunt et al. 2019; Gauci et al. 2021).

Figure 12: Observed temperature dependence of methane emission in wetland (Green), aquatic (Blue), and rice paddy (Red) environments. F is the production rate of CH_4, analogous to respiration, in Arrhenius exponential form of temperature dependence. After Durocher et al. (2014).

The popularized view of the simultaneous increase of carbon dioxide and methane, namely, that it is the result of humans, is paradoxical. Promulgated by the UN's IPCC, this view holds that their joint increase follows, purportedly, from anthropogenic sources, even though anthropogenic sources of CO_2 and CH_4 are physically independent.

In contrast, the simultaneous intensification of CO_2 emission and CH_4 emission is precisely what must accompany observed warming in the tropics. Evidenced here in observed behavior, this single physical mechanism provides a unified understanding of the joint increase of these greenhouse gases, one that follows naturally from thermally-induced emission.

In this context we have to point to an often found misinterpretation concerning the global warming potential of methane. So, the radiative efficiency of CH_4 with 3.7×10^{-4} W/m^2/ppbv is classified to be 25x larger than that of CO_2 with 1.4×10^{-5} W/m^2/ppbv. Such values are derived from the changing absorptivity or emissivity of the gases, when their concentration is changing by 1 ppbv. But these radiative efficiencies compare the two gases under completely different conditions: CH_4 at a concentration of 1.8 ppmv and CO_2 at a 200x larger concentration, when it is already strongly saturated. Also, the interference with other greenhouse gases, particularly with water vapor, is for both gases completely different.

Figure 13: a) Mean global distribution of precipitation in July. Source: GPCC (2020).
b) Mean global distribution of methane, observed by satellite. Source: ESA (2017).

Only these diverse conditions pretend a much higher radiative efficiency of methane, respectively a considerably lower value for CO_2. So, at a comparable concentration of 400

ppmv for CH_4 in the atmosphere the radiative efficiency drops to about 65% of CO_2, and at a concentration of only 1.8 ppmv of CO_2 the efficiency is about 1.5 times larger than that of CH_4 (see also Wijngaarden and Happer 2019). Only due to the different saturation strengths on quite different scales is CH_4 assumed to have a 25x larger global warming potential, while a more realistic consideration supposing a doubling each of the actual CH_4 and CO_2 concentrations shows that CH_4 does not contribute more than about 2% to global warming relative to CO_2 (Harde 2021).

6. Conclusion

Thermally-induced emission in the tropics closely tracks observed net emission of CO_2. It thus accounts for the preponderance of CO_2 net emission, which in turn determines anomalous CO_2. For this reason, the thermally-induced response to observed warming in the tropics represents nearly all of the observed increase of atmospheric CO_2.

The geographical site of net emission is noteworthy. Through the conservation law, the strong correspondence to observed net emission follows theoretically from behavior in the tropics - not in the extratropics, where anthropogenic emission is concentrated. Independently, the same correspondence to observed E_{net} follows empirically from anomalous temperature in the tropics (SH2), as well as from a time-lag analysis of anomalous CO_2 (Humlum et al. 2013). Especially noteworthy is that, in the time-lag analysis, strong coupling of anomalous CO_2 appears initially in the tropics, but then advances poleward into each hemisphere.

In both theoretical and empirical evaluations, thermally-induced emission of CO_2 represents interannual intensifications of net emission, notably during episodes of El Niño. Represented equally well is the long-term intensification of net emission during the last half century. The strong correspondence to observed changes indicates that, although operating on disparate time scales, both unsteady components of CO_2 net emission share the same physical mechanism.

Long-term intensification of E_{net} is manifest in both the observed and calculated records. It accounts for about half of the observed increase of atmospheric CO_2 (SH2). Owing to the fundamental sensitivity to temperature of surface flux (Fig. 5), such intensification is an inevitable consequence of systematic warming in the tropics.

In relation to CO_2, what is responsible for that warming is immaterial. Its influence on CO_2 should not be confused through circular reasoning. The observed warming, which forces increased CO_2 through intensified net emission, cannot itself be the result of increased CO_2. Were it, anomalous CO_2 and net emission that forces it would have increased twice as much as was observed: (i) the increase required to produce the observed warming plus (ii) the thermally-induced response to that warming, which, irrespective of what caused the warming, induces an intensification of net emission that is nearly identical to what is observed. For a detailed consideration of the "hen-or-egg causality" see also Koutsoyiannis & Kundzewicz

That net emission of CO_2 is not the cause of but the response to changes of temperature is also clear from their robust coherence, which is observed *on all time scales* (SH2). Likewise, for their in-phase relationship. Simultaneous changes of tropical temperature and net emission of CO_2 are inherent to their long-term increase as well as to interannual fluctuations, which arise unambiguously from natural changes of surface temperature. Anthropogenic emission, on the other hand, could introduce only long-term changes but not the observed interannual changes - incongruous with the strong coherence that is observed irrespective of time scale.

More fundamentally, causation by anthropogenic emission of CO_2 is eliminated by the observed *in-phase* relationship between E_{net} and tropical temperature. Simultaneous changes in those properties contradict the *in-quadrature* relationship they would have were the changes of temperature forced by changes of CO_2 (ibid). The observed in-phase relationship thus establishes unambiguously that net emission of CO_2 is forced by changes of tropical temperature - not vice

versa.

Like carbon dioxide emission, also CH_4 emission increases with temperature (Fig. 12). It too is emitted by biomass (Fig. 6), chiefly through anaerobic processes that operate in well-irrigated regions like wetlands, and those influences magnify CH_4 emission particularly from tropical land, where biomass and precipitation are abundant.

The simultaneous intensification of CO_2 emission and CH_4 emission is precisely what is expected from observed warming in the tropics. Therefore, this single physical mechanism provides a unified understanding of the joint increase of these greenhouse gases, one that follows naturally from thermally-induced emission.

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Appendix A

Observed Range of Tropical Emission

From (26) and (27), the fractional increase of atmospheric CO_2 must lie in the range:

$$\frac{\delta E_O}{E_O + 4.57\,E_O} + \frac{\delta E_L}{E_L / 0.63 + E_L} < \frac{\delta r_A}{r_A} < \frac{\delta E_O}{E_O + 0.63\,E_O} + \frac{\delta E_L}{E_L / 4.57 + E_L}, \quad \text{(A1.1)}$$

which reduces to

$$0.18\frac{\delta E_O}{E_O} + 0.39\frac{\delta E_L}{E_L} < \frac{\delta r_A}{r_A} < 0.61\frac{\delta E_O}{E_O} + 0.82\frac{\delta E_L}{E_L}. \quad \text{(A1.2)}$$

In terms of surface properties, this relationship becomes

$$0.18\frac{S_O B(v)\cdot \delta\Delta r}{S_O B(v)\cdot \Delta r} + 0.39\frac{S_L \delta R}{S_L R} < \frac{\delta r_A}{r_A} < 0.61\frac{S_O B(v)\cdot \delta\Delta r}{S_O B(v)\cdot \Delta r} + 0.82\frac{S_L \delta R}{S_L R} \quad \text{(A2.1)}$$

or
$$0.18\frac{\delta\Delta r}{\Delta r}+0.39\frac{\delta R}{R}<\frac{\delta r_A}{r_A}<0.61\frac{\delta\Delta r}{\Delta r}+0.82\frac{\delta R}{R}\,. \tag{A2.2}$$

According to (27), $E_L = S_L F_L$ is comparable to or greater than $E_O = S_O F_O$. On the other hand, coverage of the tropics by ocean exceeds that by land ($S_L < S_O$). It follows from (27) that the areal-mean flux from land must be comparable to or exceed the flux from ocean: $F_L \geq F_O$. Replacing the contribution $S_O F_O$ under exclusively ocean conditions by a greater contribution $S_L F_L$ when land is present, can only increase atmospheric back pressure (δr_A) on the ocean surface. It in turn reduces the CO_2 contrast between the atmosphere and ocean, $\delta\Delta r = \delta r_O - \delta r_A$, below its value under exclusively ocean conditions: $\delta\Delta r|_{S_L=0}$.

Consequently,
$$\frac{\delta\Delta r}{\Delta r}\leq\left.\frac{\delta\Delta r}{\Delta r}\right|_{S_L=0}\,. \tag{A3.1}$$

From (11),
$$\frac{\delta\Delta r}{\Delta r}=\left.\frac{\delta r_A}{r_A}\right|_{S_L=0}=\frac{12}{315}=0.038\,. \tag{A3.2}$$

Hence, the right-hand side of the inequality (A4.1) is already quite small. Since $\delta\Delta r/\Delta r$ must remain positive, the fractional increase of CO_2 contrast between the atmosphere and ocean must be confined to the narrow range

$$0<\frac{\delta\Delta r}{\Delta r}<\left.\frac{\delta\Delta r}{\Delta r}\right|_{S_L=0}$$

or
$$0<\frac{\delta\Delta r}{\Delta r}<0.038\,. \tag{A4.1}$$

Similarly, the fractional increase in respiration follows from (22) - (24) as

$$\frac{\delta R}{R}=0.322\cdot 1.2=0.386\,. \tag{A4.2}$$

With (A4), (A2) reduces to

$$0.18\cdot 0+0.386\cdot 0.386<\frac{\delta r_A}{r_A}<0.614\cdot 0.038+0.82\cdot 0.386\,,$$

or finally
$$0.15<\frac{\delta r_A}{r_A}<0.34\,. \tag{A5}$$

Appendix B

Environment as a Net Sink or Net Emitter?

According to the conservation law (1) changes of the CO_2 mixing ratio r_A are controlled by a competition between the total emission rate E and its removal through an absorption rate A:

$$\frac{dr_A}{dt}=E-A=E-\frac{r_A}{\tau_{eff}}\,, \tag{B1}$$

where the absorption is proportional to the instantaneous abundance of CO_2 and τ_{eff} is the effective absorption time.

With a constant emission E_0 and a perturbation ΔE, Eq. (B1) can also be expressed in the form:

$$\frac{dr_A}{dt} - \Delta E = E_0 - \frac{r_A}{\tau_{eff}} . \tag{B2}$$

For ΔE as the anthropogenic emission rate, as derived from the CDIAC-data, and with the observed increase of the mixing ratio $\Delta r_A/\Delta t$ over one year at Mauna Loa it is obvious that the net flux, E_0 - A can quite well be assessed without needing to know the absolute magnitudes of E_0 or A, quantities which on their parts are often highly uncertain. It is also evident that in a linearly responding system without some virtual amplification the changes $\Delta r_A/\Delta t$ cannot be greater than the perturbation itself, and that these changes are responding with some delay to the emissions, controlled by the effective absorption time τ_{eff}. So, with the left side of (B2) negative, also the right side is negative, and the environmental uptake $A = r_A/\tau_{eff}$ must be larger than the natural emissions E_0. From this statement, however, often wrong conclusions are derived that the environment must have acted as a net sink throughout the Industrial Era, and thus, nature could not have been the reason for any observed CO_2 increase.

For simplicity considering for a moment a constant perturbation ΔE, native or anthropogenic origin or both, the general solution of (B2) is:

$$r_A(t) = r_A(0) \cdot e^{-t/\tau_{eff}} + (E_0 + \Delta E) \cdot \tau_{eff} \cdot \left(1 - e^{-t/\tau_{eff}}\right), \tag{B3}$$

and with $r_A(0) = E_0 \cdot \tau_{eff}$ as the equilibrium concentration for the constant emission E_0, the excess concentration $\Delta r_A(t) = r_A(t) - r_A(0)$ becomes

$$\Delta r_A(t) = \Delta E \cdot \tau_{eff} \cdot \left(1 - e^{-t/\tau_{eff}}\right). \tag{B4}$$

Assuming an increase of the mixing ratio over one year, as derived from an airborne fraction of AF = 44%, but this growth related to the maximum expected increase at a new equilibrium $\Delta r_A^{eq} = \Delta E \cdot \tau_{eff}$, gives a conditional equation with an upper limit for τ_{eff}:

$$\begin{aligned} 0.44 \cdot \Delta r_A^{eq} &= \Delta E \cdot \tau_{eff} \cdot \left(1 - e^{-1\,yr/\tau_{eff}}\right), \\ 0.44 &= 1 - e^{-1\,yr/\tau_{eff}} \end{aligned} \tag{B5.1}$$

or

$$\tau_{eff} = -\frac{1\,yr}{ln(1 - 0.44)} = 1.72 \text{ yrs} . \tag{B5.2}$$

Relating the 44% increase directly to the emission rate ΔE, also over one year (Δt = 1 yr), in agreement with the definition of the airborne fraction, this gives:

$$\begin{aligned} 0.44 \cdot \Delta E \cdot \Delta t &= \Delta E \cdot \tau_{eff} \cdot \left(1 - e^{-\Delta t/\tau_{eff}}\right) \\ 0.44 \cdot \Delta t &= \tau_{eff} \cdot \left(1 - e^{-\Delta t/\tau_{eff}}\right) \end{aligned} \tag{B5.3}$$

and can be seen as a condition for a lower limit, yielding an effective absorption time of only τ_{eff} = 0.51 yrs, close to the calculations for thermally induced emissions of interannual fluctuations with τ_{eff} = 0.67 yrs (Eq.(51.2)). These rough estimates confirm quite well the independently derived result for anthropogenic absorption (51.3) with τ_{eff} = 1.1 yrs, and they explain, why the airborne fraction is largely independent of the size of a perturbation ΔE.

The preceding considerations make also clear that from a negative left hand side of (B2) or a lower concentration $\Delta r_A(\Delta t) \leq \Delta E \cdot \Delta t$ (see Eqs.(B5)) we cannot simply infer that nature is a net sink and therefore could not be responsible for any increasing CO_2 level.

Related to a constant native emission E_0 and the initial equilibrium mixing ratio $r_A(0) = E_0 \cdot \tau_{eff}$, nature surely acts as a further increasing sink, driven by the additional native or human emissions ΔE. Thus, the environment indeed looks like a net absorber. But this absorption tracks the anomalous emission ΔE with some delay and cannot be larger than this emission, which in reali-

ty is even further increasing over time. Because native emission and absorption processes are largely independent phenomena, also with an increasing absorption can nature still be a net emitter to the atmosphere and even by far the strongest donator, as this and our previous studies (SH2; Harde 2019, Figs 8-10, Subsec. 5.3) clearly demonstrate.

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