Radiative Forcing and Feedbacks in Climate Change

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Definitions

Climate system:

(atmosphere + hydrosphere + lithosphere + biosphere)

+ mutual interactions

+ responses to external influences = Earth's climate

Climate forcing:

energy imbalance imposed on the climate system

<u>Direct radiative forcing</u>:

forcing that directly affects Earth's radiative budget

(e.g.: added CO_2 absorbs and emits infrared radiation)

<u>Indirect radiative forcing</u>:

forcing that creates radiative imbalance by first altering climate system components (e.g.: added aerosols modify cloud properties)

<u>Nonradiative forcing</u>:

forcing creates energy imbalance not directly involving radiation

(e.g.: agricultural irrigation increases evapotranspiration flux)

Climate response:

change in climate system resulting from a climate forcing

Climate feedback:

amplification or dampening of response due to changes in system

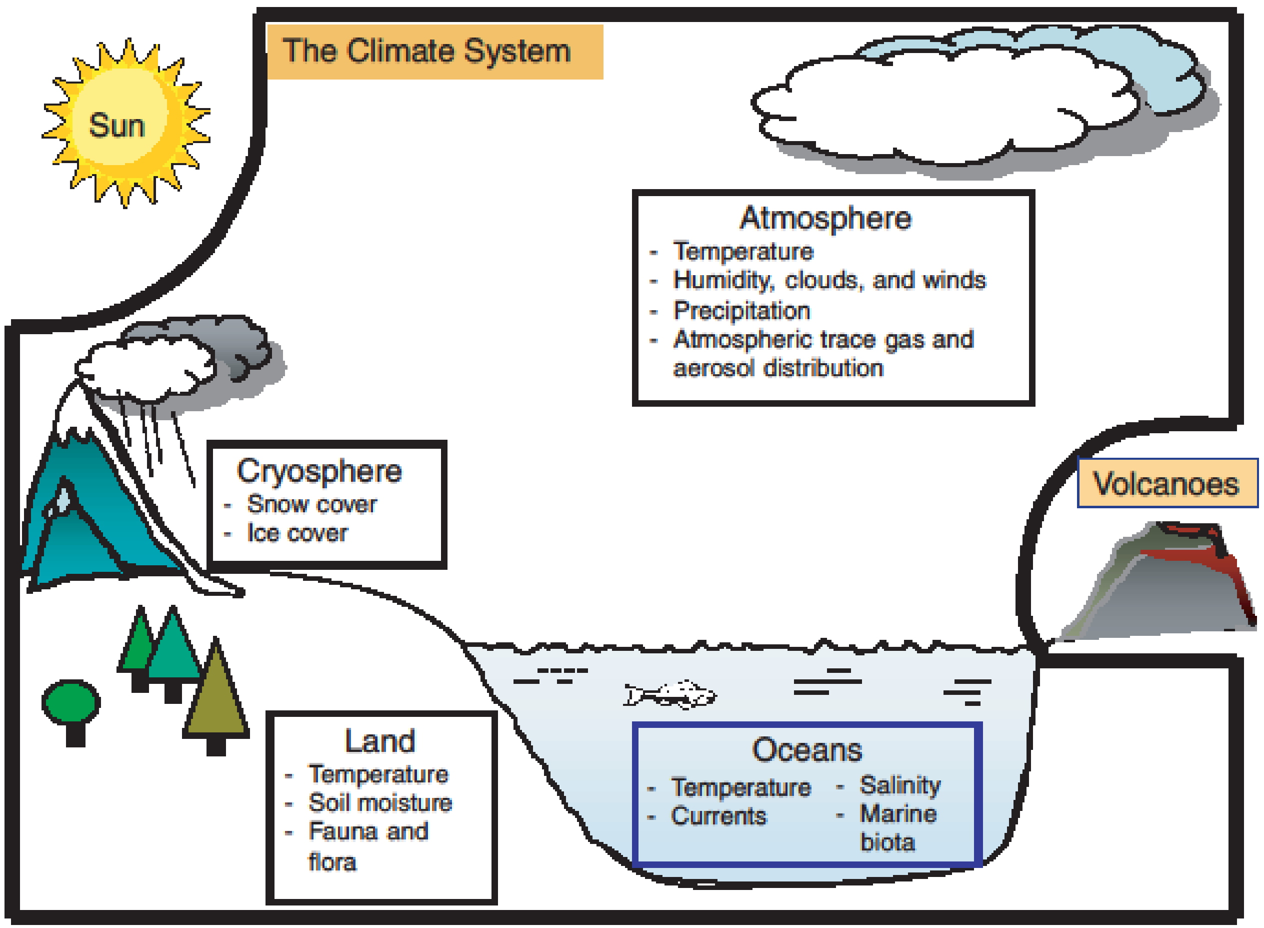


FIGURE 1-1 The climate system, consisting of the atmosphere, oceans, land, and cryosphere. Important state variables for each sphere of the climate system are listed in the boxes. For the purposes of this report, the Sun, volcanic emissions, and human-caused emissions of greenhouse gases and changes to the land surface are considered external to the climate system.

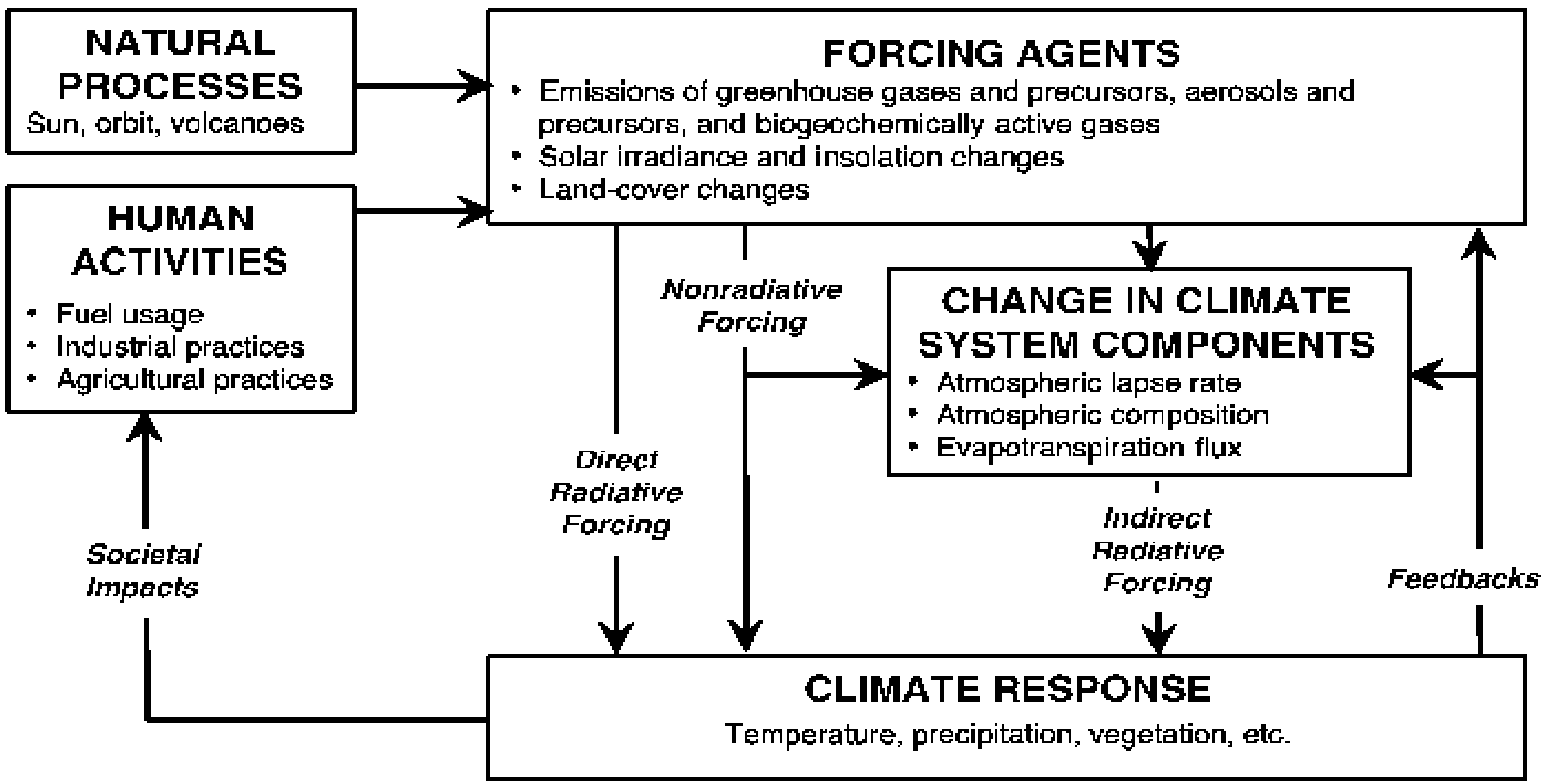
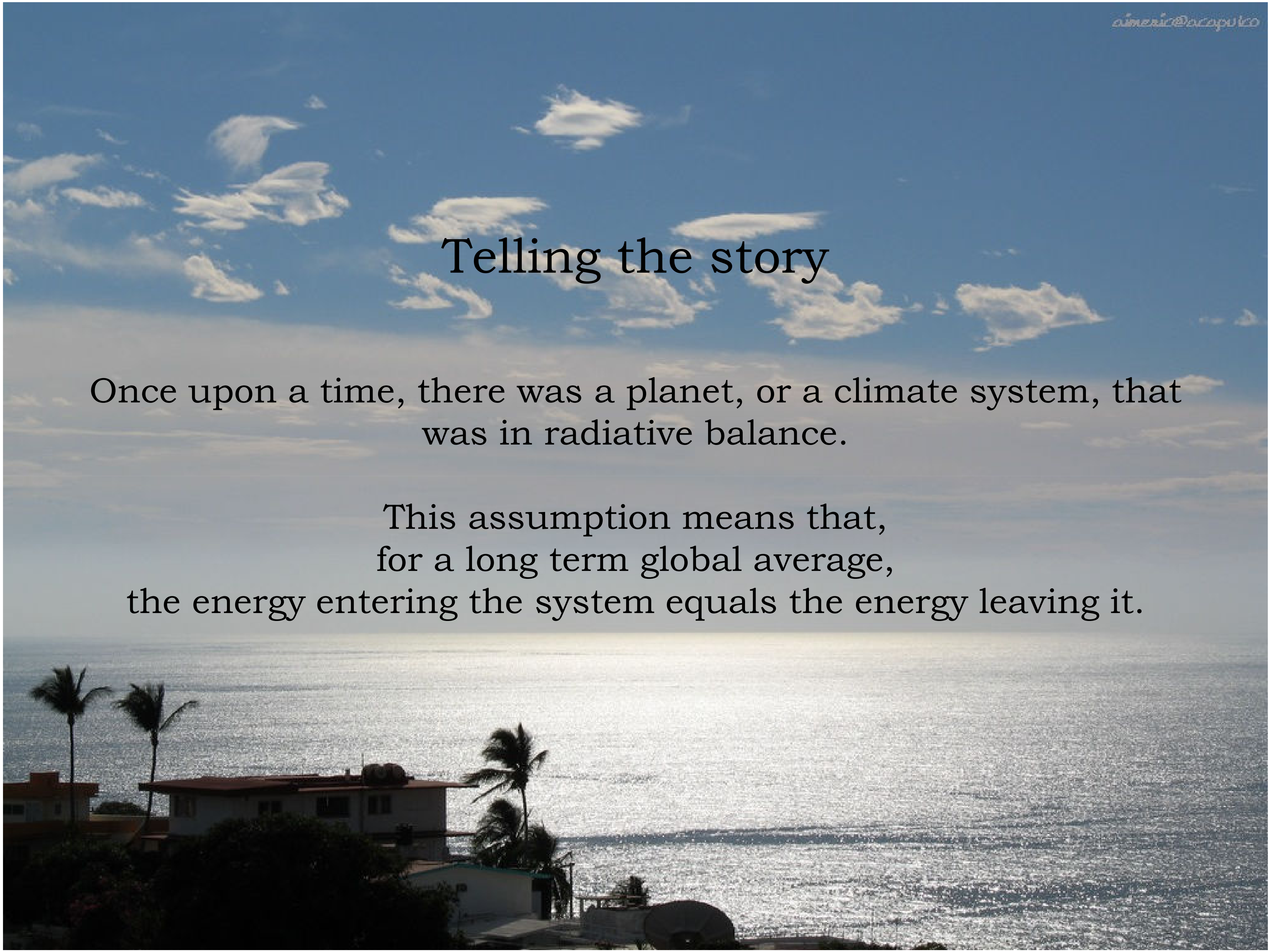


FIGURE 1-2 Conceptual framework of climate forcing, response, and feedbacks under present-day climate conditions. Examples of human activities, forcing agents, climate system components, and variables that can be involved in climate response are provided in the lists in each box.



Telling the story

Once upon a time, there was a planet, or a climate system, that was in radiative balance.

This assumption means that,
for a long term global average,
the energy entering the system equals the energy leaving it.

107 + 235 = 342

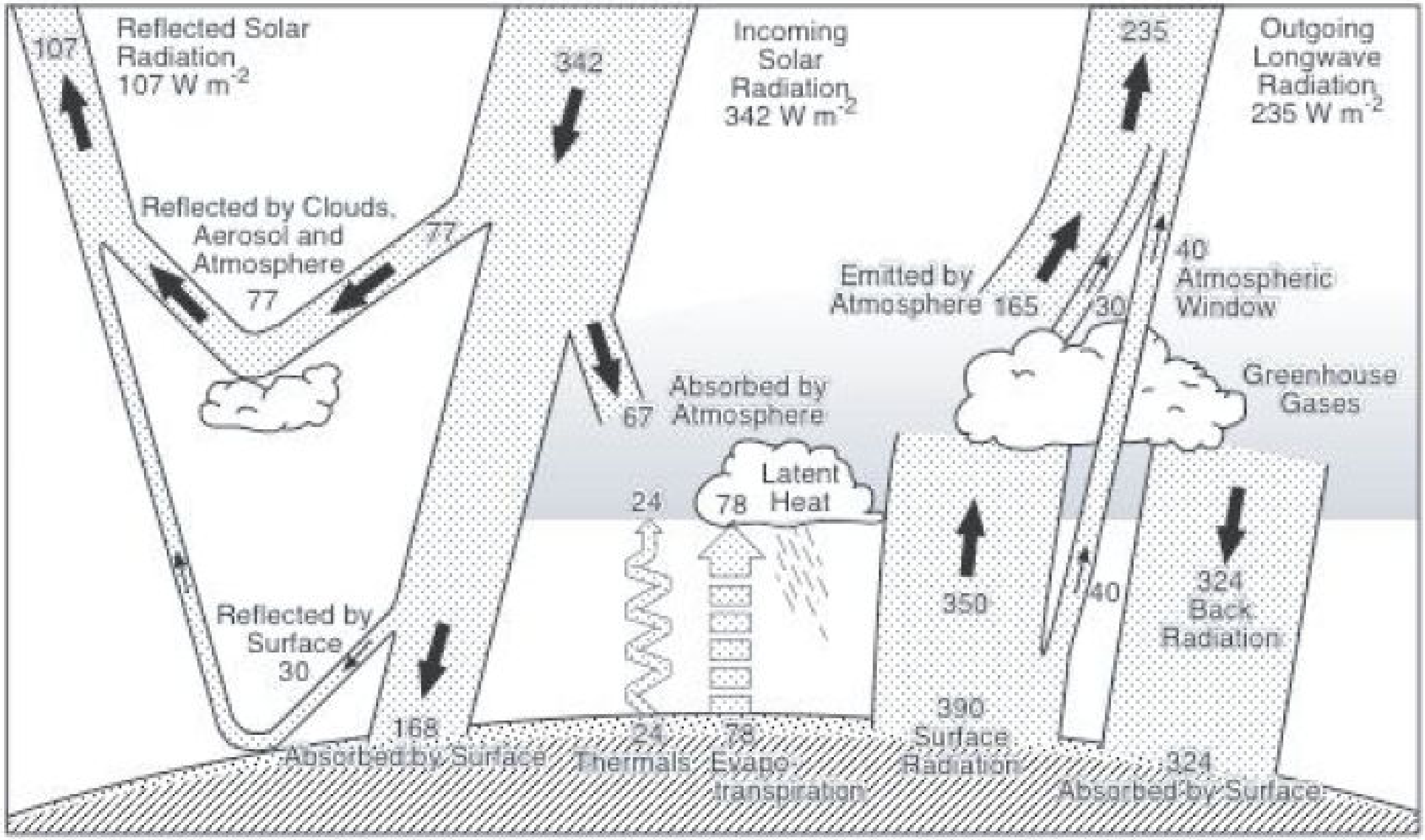


FIGURE 1-3 Energy budget for the atmospheric components of the climate system.
SOURCE: Kiehl and Trenberth (1997).

(Global annual mean for the period: 1985 to 1989).

102 + 239 = 341

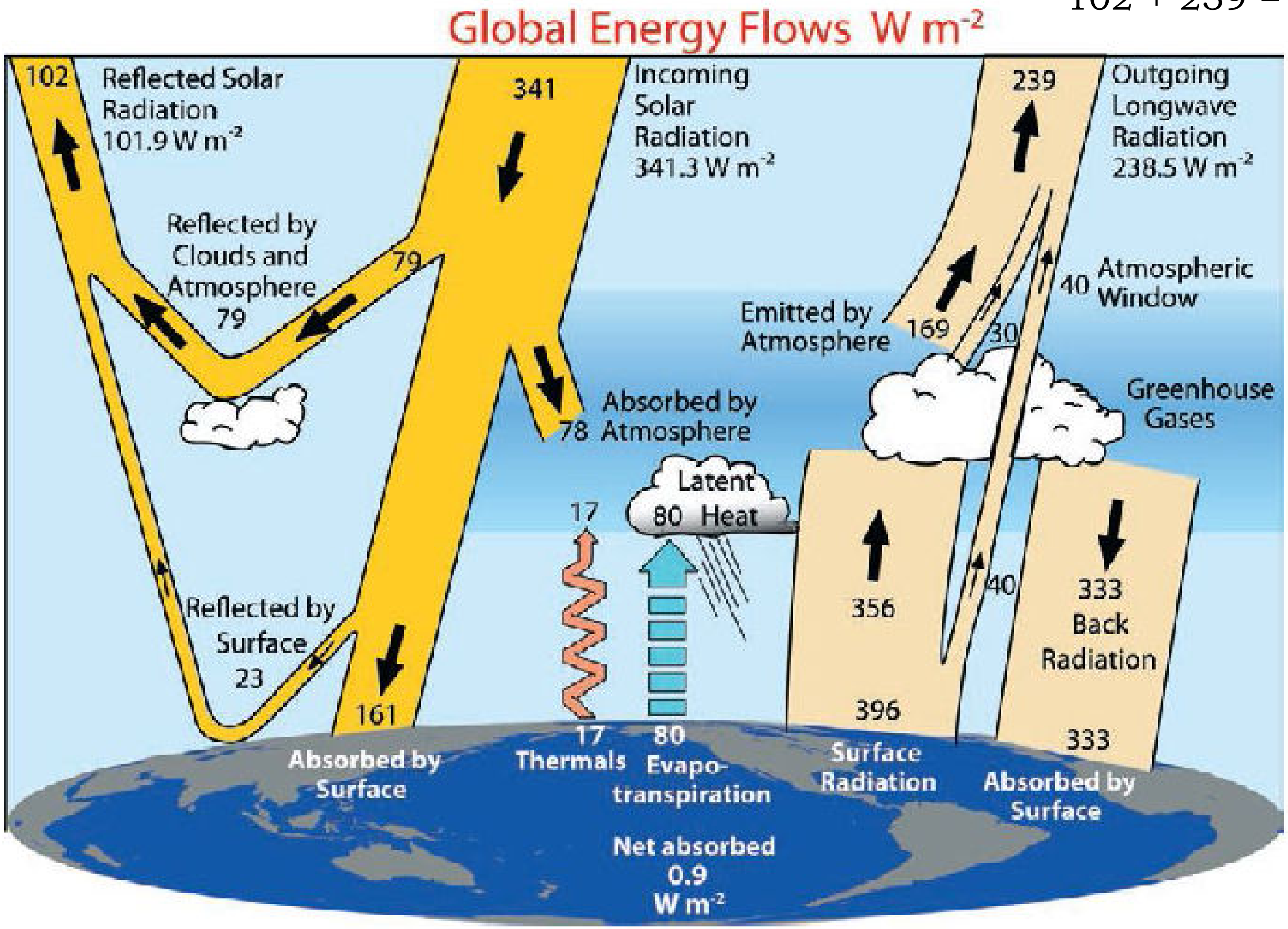
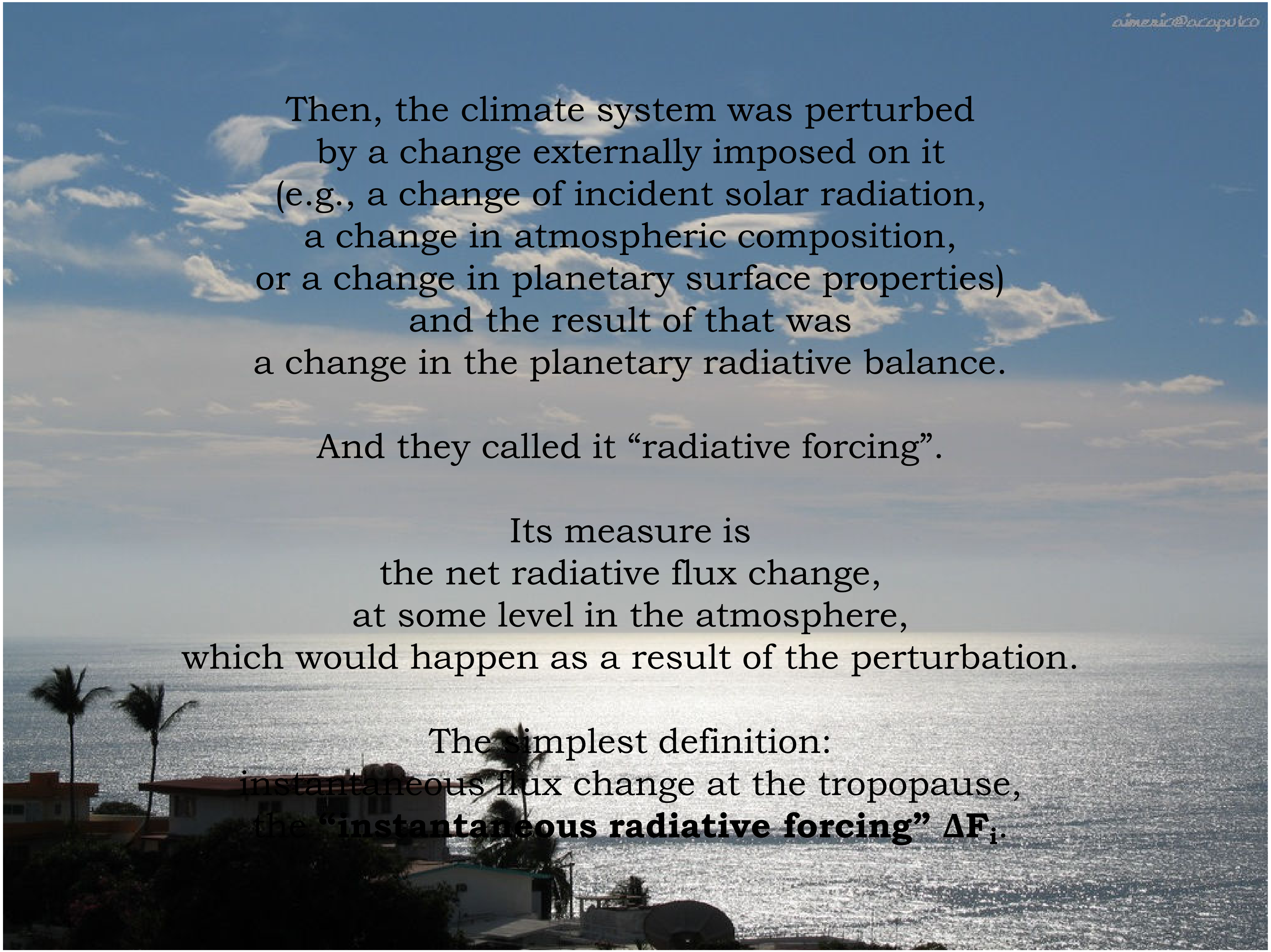


Fig. 1. The global annual mean Earth's energy budget for the Mar 2000 to May 2004 period (W m^{-2}). The broad arrows indicate the schematic flow of energy in proportion to their importance. **(Trenberth et al., 2009).**

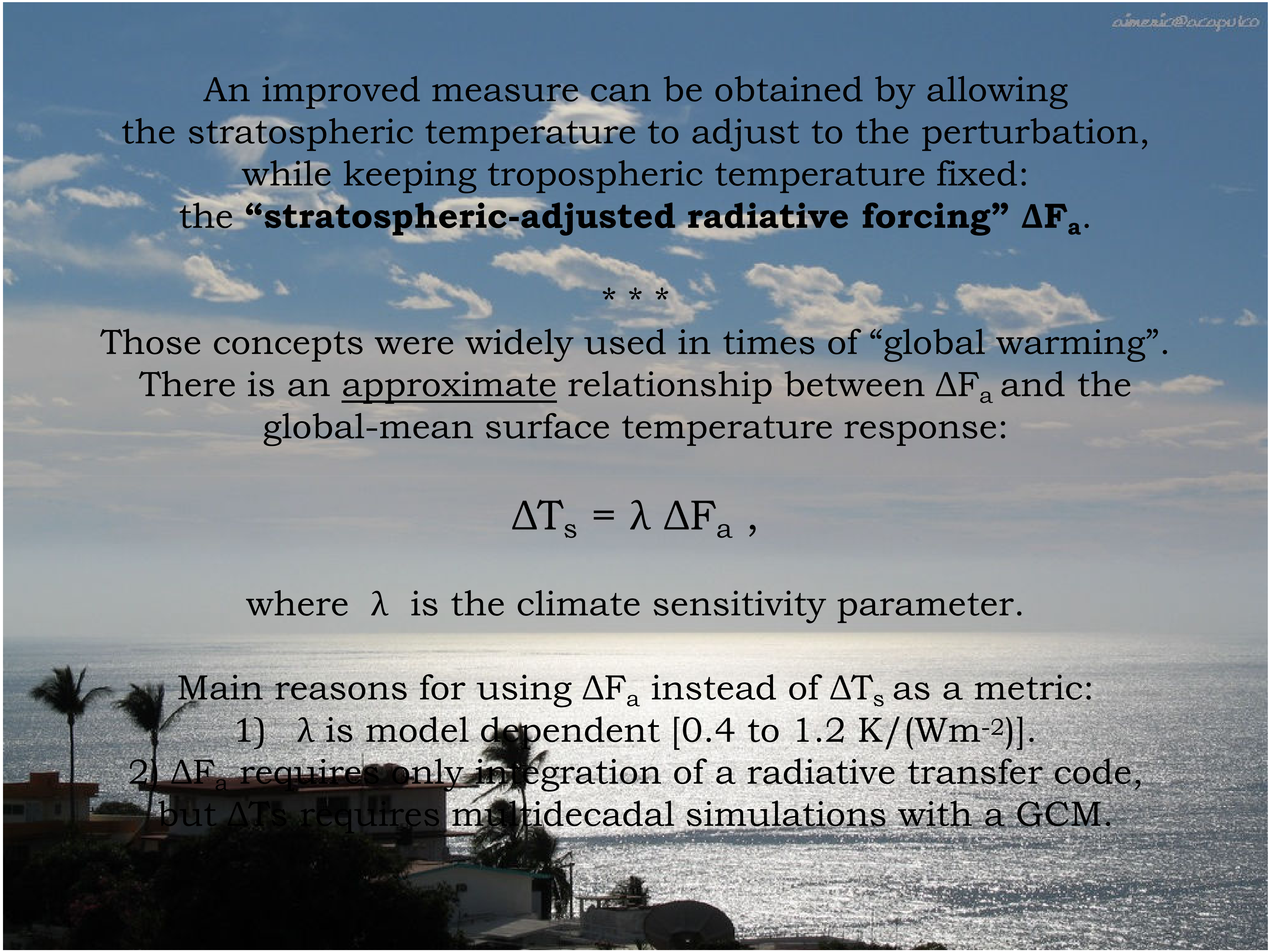


Then, the climate system was perturbed
by a change externally imposed on it
(e.g., a change of incident solar radiation,
a change in atmospheric composition,
or a change in planetary surface properties)
and the result of that was
a change in the planetary radiative balance.

And they called it “radiative forcing”.

Its measure is
the net radiative flux change,
at some level in the atmosphere,
which would happen as a result of the perturbation.

The simplest definition:
instantaneous flux change at the tropopause,
the **“instantaneous radiative forcing”** $\mathbf{\Delta F_i}$.



An improved measure can be obtained by allowing the stratospheric temperature to adjust to the perturbation, while keeping tropospheric temperature fixed: the **“stratospheric-adjusted radiative forcing” ΔF_a**.

* * *

Those concepts were widely used in times of “global warming”. There is an approximate relationship between ΔF_a and the global-mean surface temperature response:

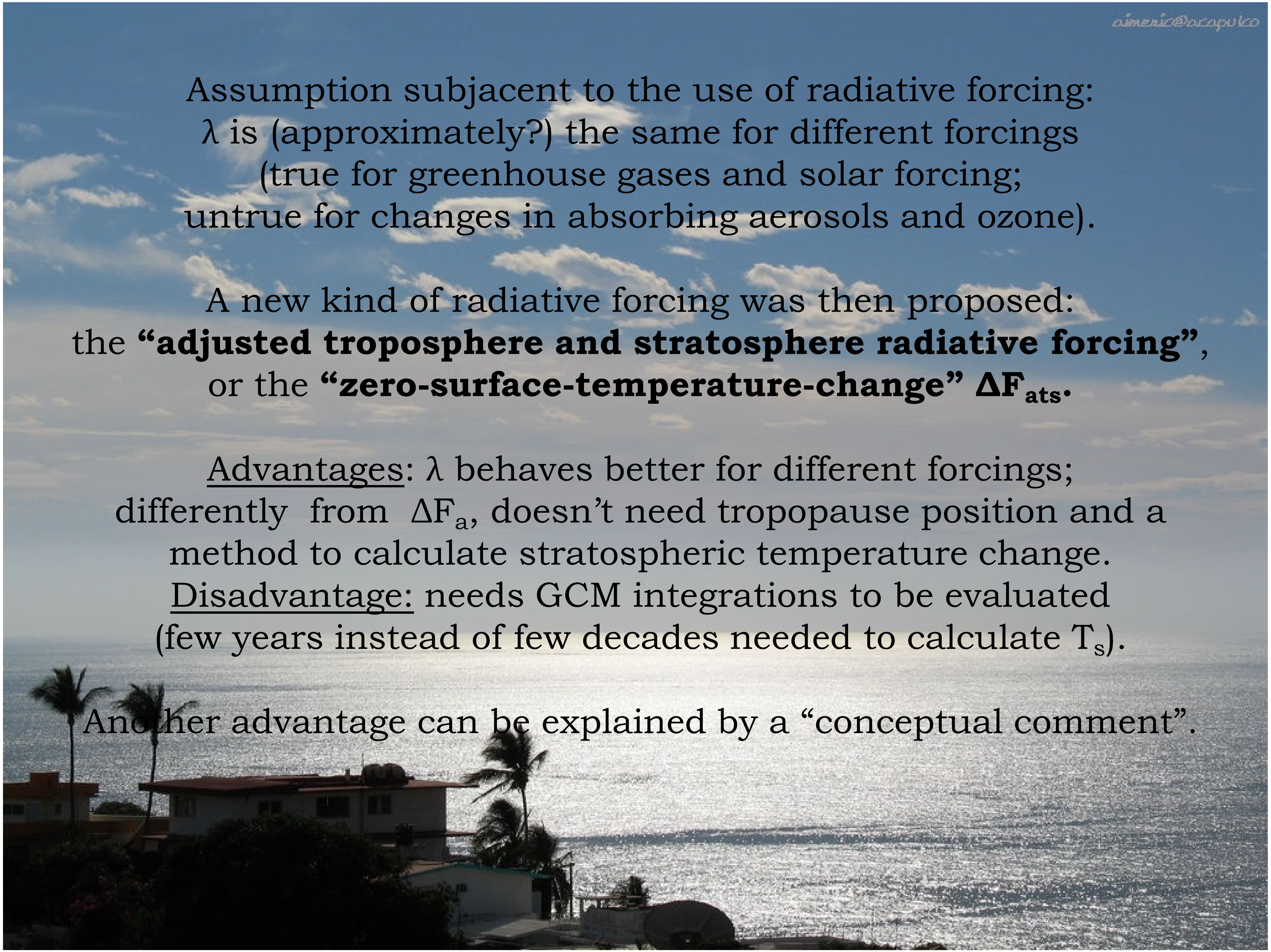
$$\Delta T_s = \lambda \, \Delta F_a \, ,$$

where λ is the climate sensitivity parameter.

Main reasons for using ΔF_a instead of ΔT_s as a metric:

1) λ is model dependent [0.4 to 1.2 K/(Wm^{-2})].

2) ΔF_a requires only integration of a radiative transfer code, but ΔT_s requires multidecadal simulations with a GCM.

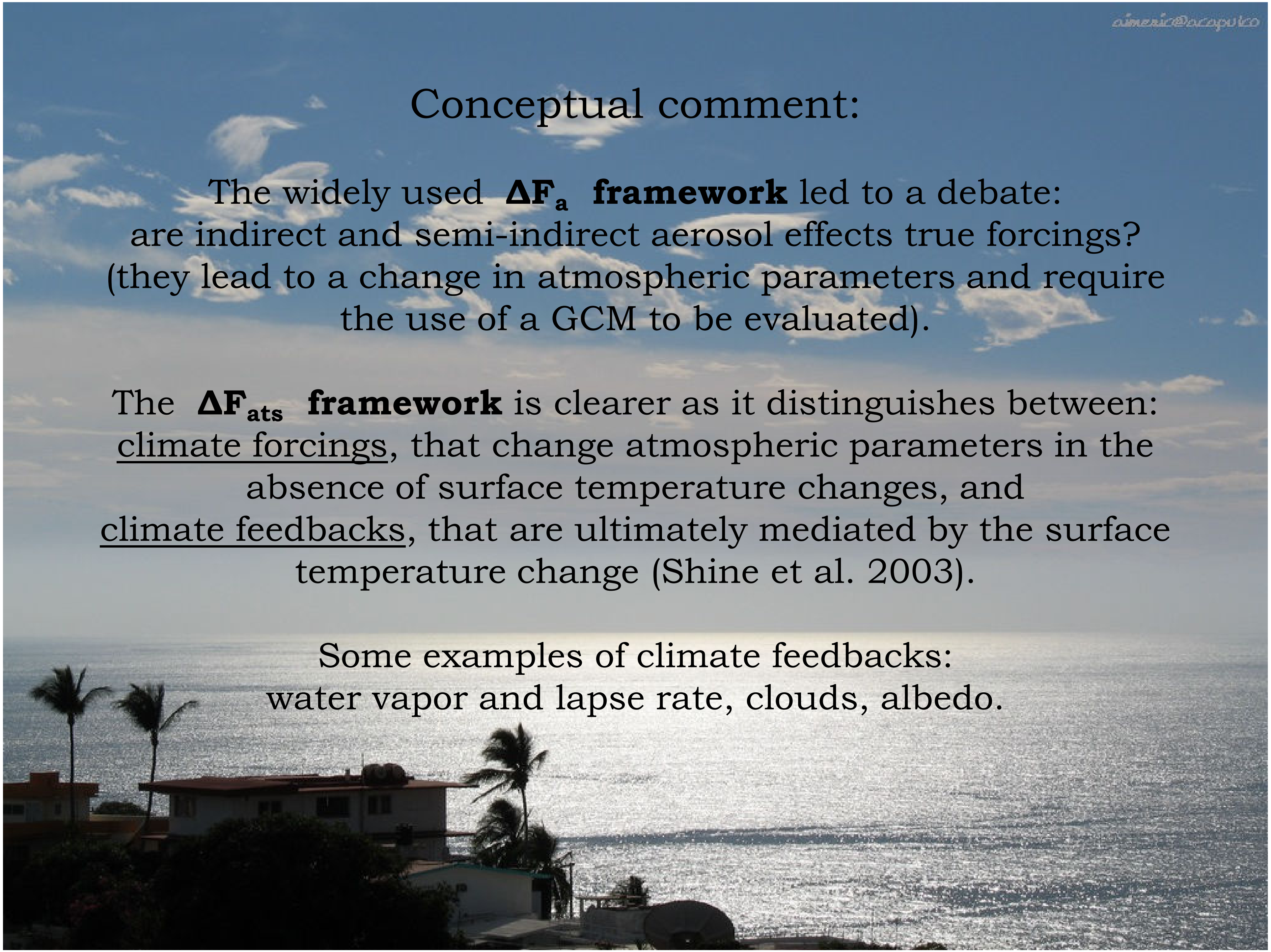


Assumption subjacent to the use of radiative forcing:
λ is (approximately?) the same for different forcings
(true for greenhouse gases and solar forcing;
untrue for changes in absorbing aerosols and ozone).

A new kind of radiative forcing was then proposed:
the **"adjusted troposphere and stratosphere radiative forcing"**,
or the **"zero-surface-temperature-change" ΔF_{ats}.**

<u>Advantages</u>: λ behaves better for different forcings;
differently from ΔF_a, doesn't need tropopause position and a
method to calculate stratospheric temperature change.
<u>Disadvantage:</u> needs GCM integrations to be evaluated
(few years instead of few decades needed to calculate T_s).

Another advantage can be explained by a "conceptual comment".



Conceptual comment:

The widely used **ΔF_a framework** led to a debate:
are indirect and semi-indirect aerosol effects true forcings?
(they lead to a change in atmospheric parameters and require the use of a GCM to be evaluated).

The **ΔF_{ats} framework** is clearer as it distinguishes between:
climate forcings, that change atmospheric parameters in the absence of surface temperature changes, and
climate feedbacks, that are ultimately mediated by the surface temperature change (Shine et al. 2003).

Some examples of climate feedbacks:
water vapor and lapse rate, clouds, albedo.

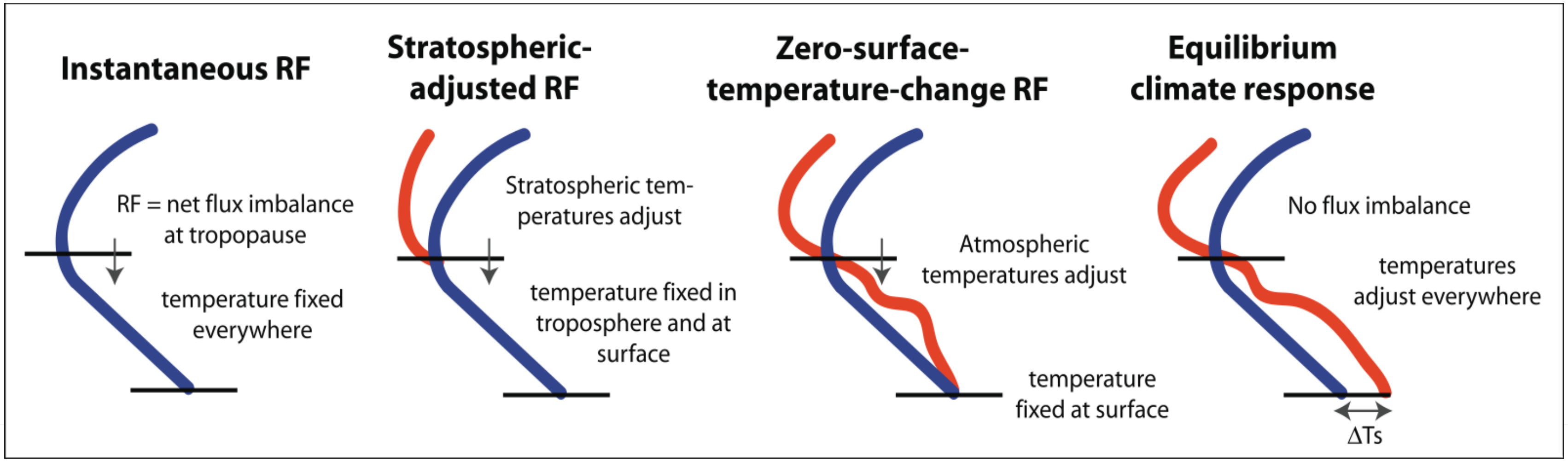


Figure 2.2. *Schematic comparing* ***RF calculation methodologies****. Radiative forcing, defined as the net flux imbalance at the tropopause, is shown by an arrow. The horizontal lines represent the surface (lower line) and tropopause (upper line). The unperturbed temperature profile is shown as the blue line and the perturbed temperature profile as the red line. From left to right:* ***Instantaneous RF****: atmospheric temperatures are fixed everywhere;* ***stratospheric-adjusted RF****: allows stratospheric temperatures to adjust;* ***zero-surface-temperature-change RF****: allows atmospheric temperatures to adjust everywhere with surface temperatures fixed; and* ***equilibrium climate response****: allows the atmospheric and surface temperatures to adjust to reach equilibrium (no tropopause flux imbalance), giving a surface temperature change (ΔT_s).*

(Forster et al. 2007, chapter 2 of the IPCC Report)

Radiative forcing

… is a change imposed on the planetary radiation balance (…) measured by the net radiative flux, at some level in the atmosphere, calculated to occur in response to the perturbation, which may be a change of incident solar radiation, atmospheric composition, or planetary surface properties, for example (Hansen et al. 1997).

… is defined as the change in the irradiance at the tropopause following, for example, an increase in carbon dioxide concentration or a change in solar output (Shine et al. 2003).

… has traditionally been defined as a change in energy flux at the tropopause resulting from a change in a component external to the physical climate system (National Research Council, 2005).

… is the change in net (down minus up) irradiance (solar plus longwave; in W m^{-2}) at the tropopause after allowing for stratospheric temperatures to readjust to radiative equilibrium, but with surface and tropospheric temperatures and state held fixed at the unperturbed values (Forster et al. 2007, the IPCC definition).

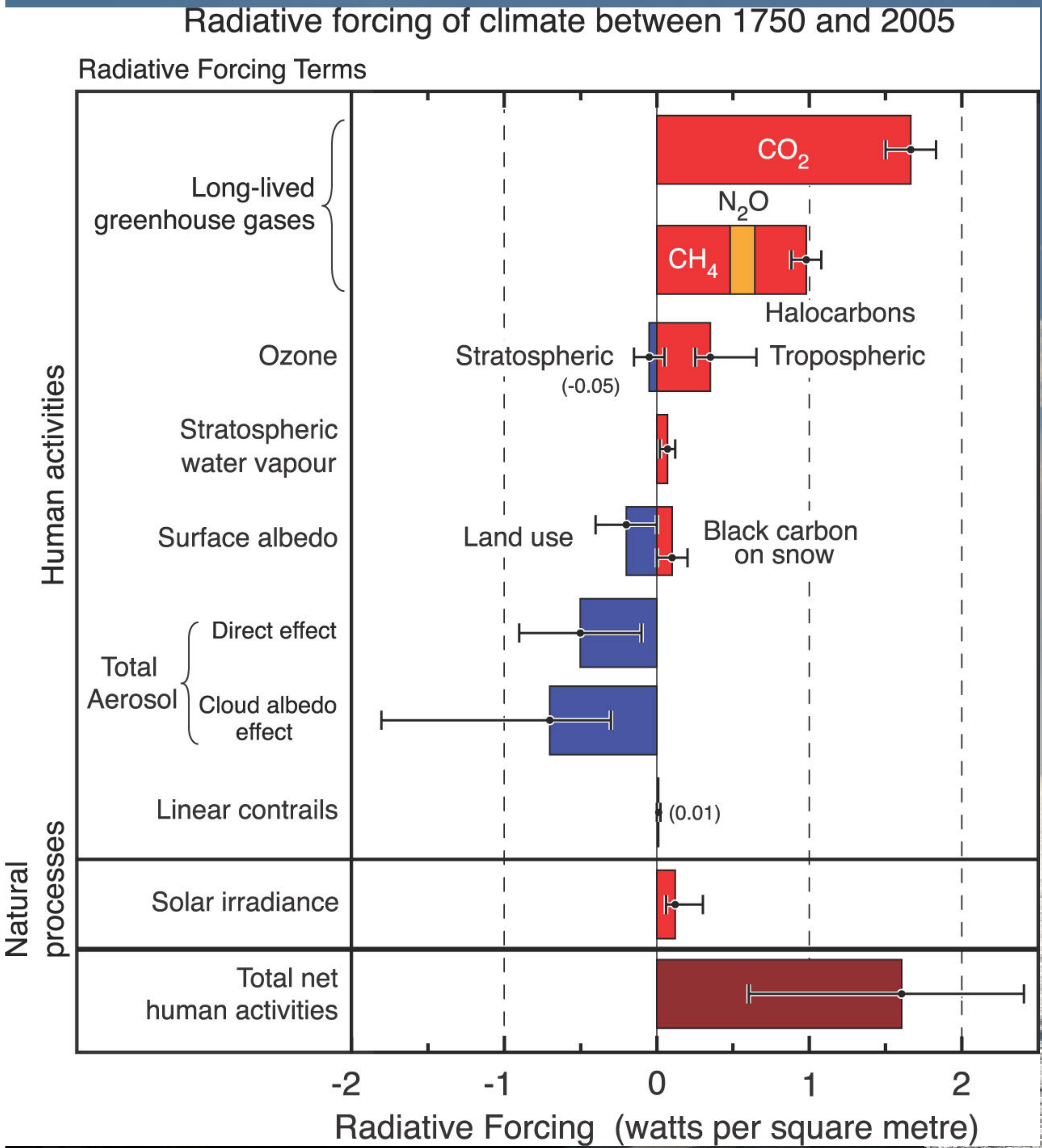


Figure 2. Summary of the principal components of the radiative forcing of climate change.

All radiative forcings result from factors that affect climate and are associated with human activities or natural processes.

The values represent forcings in 2005 relative to the start of the industrial era (about 1750).

Human activities cause significant changes in long-lived gases, ozone, water vapor, surface albedo, aerosols and contrails.

The only increase in natural forcing of any significance between 1750 and 2005 occurred in solar irradiance.

Positive forcings lead to warming of climate and negative forcings lead to cooling. The thin black line attached to each colored bar represents the range of uncertainty for the respective value.

(IPCC, 2007).

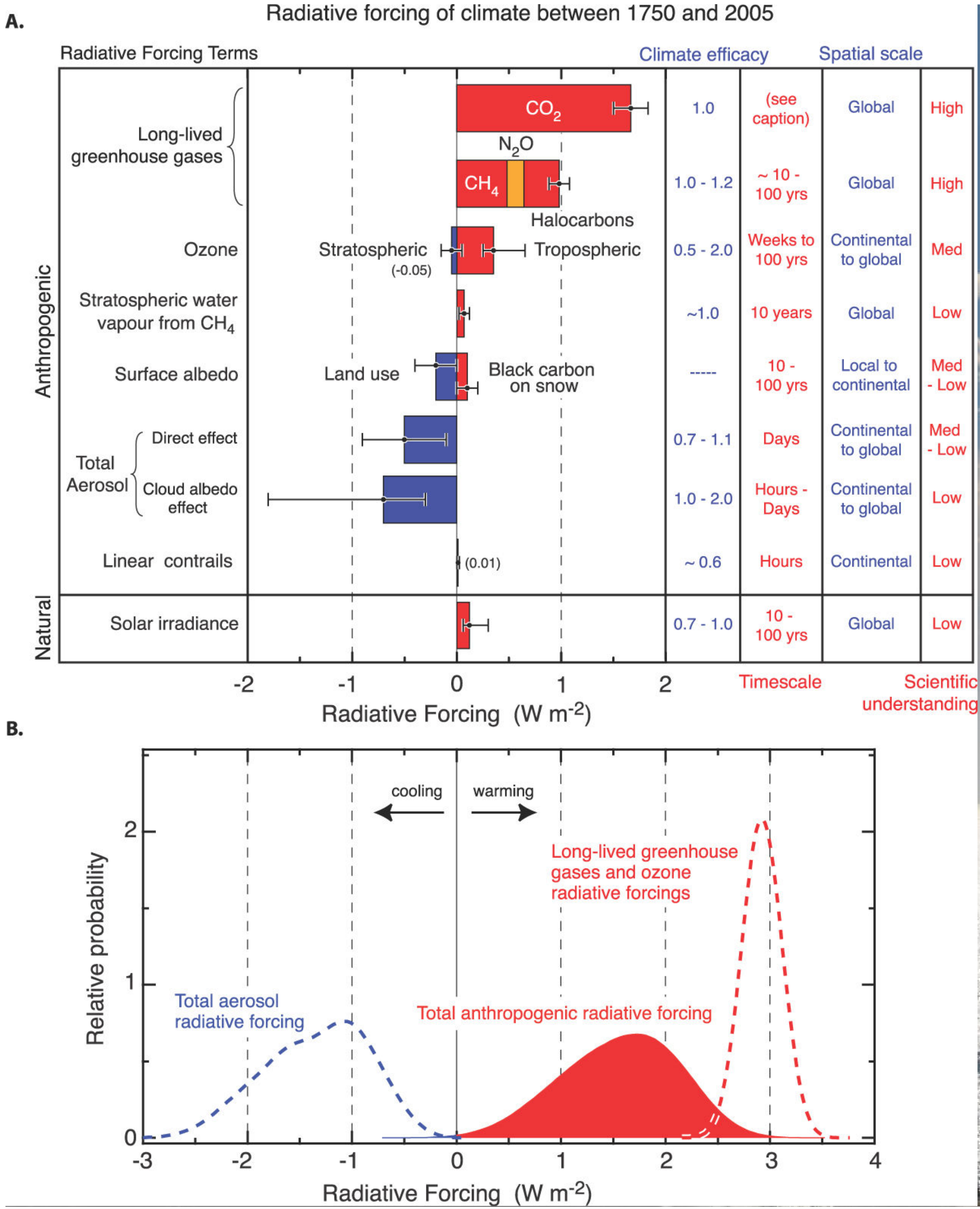


Figure 2.20. ***(A)*** *Global mean RFs from the agents and mechanisms discussed in this chapter, grouped by agent type. Anthropogenic RFs and the natural direct solar RF are shown. The plotted RF values correspond to the bold values in Table 2.12. Columns indicate other characteristics of the RF; efficacies are not used to modify the RFs shown. Time scales represent the length of time that a given RF term would persist in the atmosphere after the associated emissions and changes ceased. No CO2 time scale is given, as its removal from the atmosphere involves a range of processes that can span long time scales, and thus cannot be expressed accurately with a narrow range of lifetime values. The scientific understanding shown for each term is described in Table 2.11.*

(B) *Probability distribution functions (PDFs) from combining anthropogenic RFs in (A). Three cases are shown: the total of all anthropogenic RF terms (block filled red curve; see also Table 2.12); LLGHGs and ozone RFs only (dashed red curve); and aerosol direct and cloud albedo RFs only (dashed blue curve). Surface albedo, contrails and stratospheric water vapour RFs are included in the total curve but not in the others. For all of the contributing forcing agents, the uncertainty is assumed to be represented by a normal distribution (and 90% confidence intervals) with the following exceptions: contrails, for which a lognormal distribution is assumed to account for the fact that the uncertainty is quoted as a factor of three; and tropospheric ozone, the direct aerosol RF (sulfate, fossil fuel organic and black carbon, biomass burning aerosols) and the cloud albedo RF, for which discrete values based on Figure 2.9, Table 2.6 and Table 2.7 are randomly sampled. Additional normal distributions are included in the direct aerosol effect for nitrate and mineral dust, as these are not explicitly accounted for in Table 2.6. A one-million point Monte Carlo simulation was performed to derive the PDFs (Boucher and Haywood, 2001). Natural RFs (solar and volcanic) are not included in these three PDFs. Climate efficacies are not accounted for in forming the PDFs.*

(IPCC, 2007).

Other concepts and ways of calculating radiative forcing
(when "global warming" become "climate change")

Global mean radiative forcing at the surface:

A forcing agent
(e.g. absorbing aerosols, land-use changes)
alters the heating profile and makes ΔF_a or ΔF_{ats} not directly related to T_s.
Combining global mean radiative forcing at the surface and at TOA could resolve this limitation. Could also assess responses beyond surface temperature change.

Regional radiative forcing:

Forcings with great spatial variability can have regional magnitudes much greater than their global averages.
Regional radiative forcing provide better measure of regional response.

Regional nonradiative forcing:

Some forcings affect climate in nonradiative ways, by modifying hydrological cycle or vegetation dynamics.

Ocean heat content:

Change in ocean heat storage with time can be used to calculate the net radiative imbalance of the Earth.
However, there are several open questions on the accuracy of calculations of ocean heat content.

In the following a summary of a table from National Research Council (2005).

Climate forcing metric	Strengths	Limitations
Stratospheric-adjusted radiative forcing at tropopause ΔF_a	• ΔT_s nearly linearly related to ΔF_a • computationally efficient • enables comparison of different forcing agents • enables comparison of different models, benchmarks and literature • can be used in simple models for policy analysis • already in policy dialogue	• insufficient information on hydrological response • doesn't characterize absorbing aerosols impact • doesn't characterize regional response • doesn't accommodate nonlinear response • doesn't characterize impact of nonradiative forcing, and of indirect and semidirect aerosol effect
Adjusted troposphere and stratosphere radiative forcing	• characterizes indirect and semidirect aerosol effect • incorporates fast feedbacks in simulation of climate forcing and response • insensitive to altitude where forcing is calculated	• except for indirect and semidirect aerosol effect, same limitations of ΔF_a • not computationally efficient • not comparable across models • doesn't characterize regional structure

Climate forcing metric	Strengths	Limitations
Global mean radiative forcing at surface	• characterizes the surface energy budget • reported with TOA forcing, may provide information on lapse rate effects	• doesn’t characterize regional structure
Regional radiative forcing (direct and indirect)	• may provide better measure of regional climate response • allows characterization of teleconnected reponse	• work needed to quantify links of regional forcing to regional and global climate response
Regional nonradiative forcing (hydrological, land, biogeochemical)	• recognizes additional nonradiative forcings • allows characterization of teleconnected reponse • provides more complete characterization of radiative forcing	• no widely accepted metrics • work needed to quantify links of regional forcing to regional and global climate response • some types not easily in watts per square meter
Ocean heat content	• can be used to calculate net radiative imbalance of the Earth • constraint on performance of climate models	• frequency and coverage of observations may be insufficient to accurate determination of radiative imbalance

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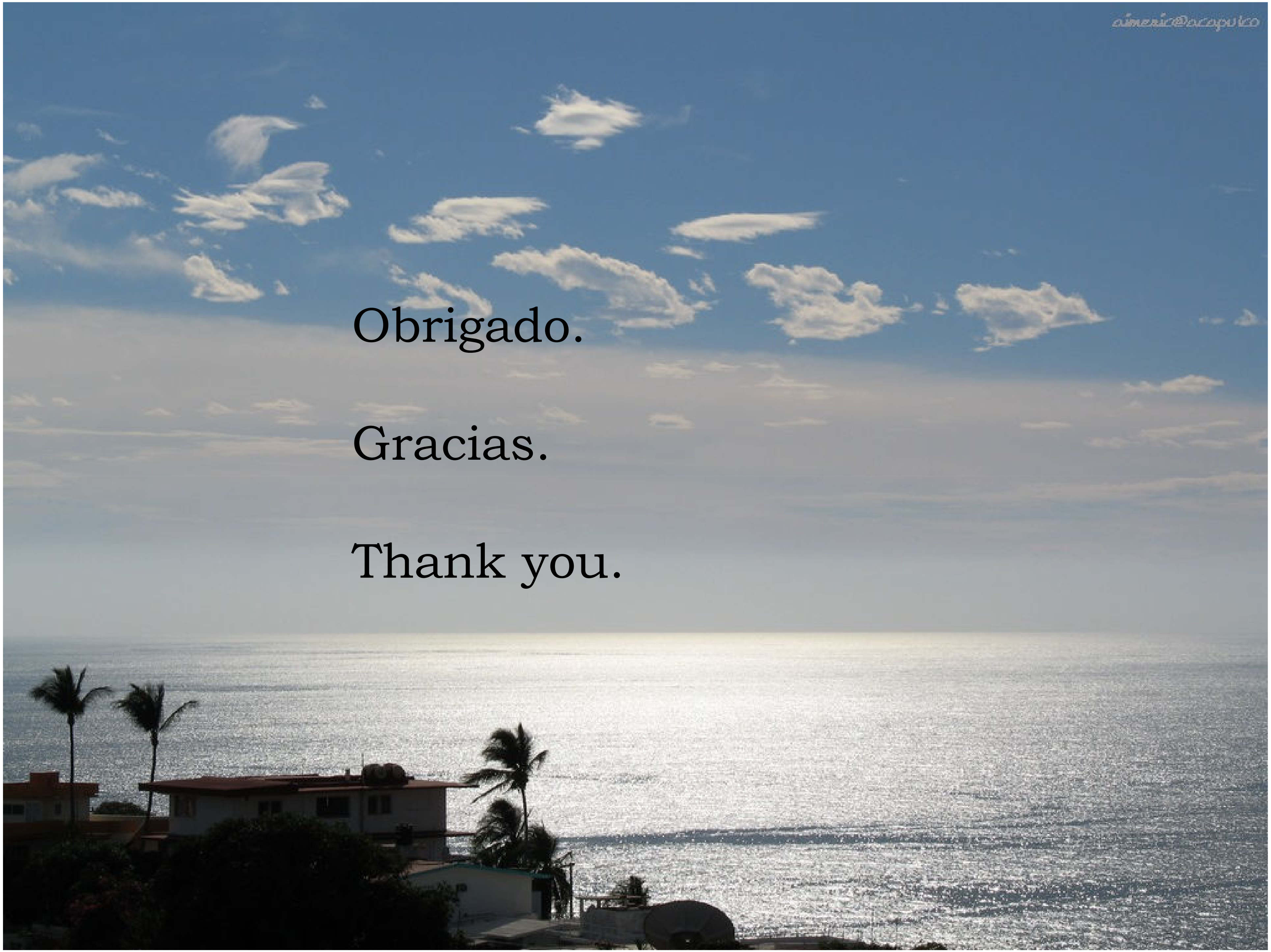

aimeric@acapulco
Obrigado.
Gracias.
Thank you.