Twelve Plus One Lectures: A climate system view of clouds and convection

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Contents

1 Energy balance description of the Earth System

Earth is a non-equilibirum system that exchanges energy with the external environment (stars and space) through fluxes of radiant energy, irradiance. The broadband irradiance incident at the top of the atmosphere is given by the total solar irradiance, Q_0. Because the solar irradiance is predominantly concentrated in the short, or visible wavelengths, it is sometimes call the shortwave, or visible irradiance, in contrast to the longer wavelenths that dominate the irradiance that originates from the terrestrial source. As a corollary the latter is often called the longwave irradiance,.

For the solar minimum in 2008, Q_0, is estimated to have been $1360.8 \pm 0.5\,\mathrm{W\,m^{-2}}$, and this is usually assumed to be a constant baseline value – giving rise to the idea of the *solar constant*. However the solar irradiance varies on a variety of timescales, notably with the solar (sunspot) cycle whose period is roughly eleven years and over which the irradiance may increase by 1-2 $\mathrm{W\,m^{-2}}$. But the solar irradiance also varies across solar minima, hence the concept of a constant sun can be misleading and the term total solar irradiance, TSI, is preferred to the somewhat older concept of the solar constant. For the purposes of studying changes to the energy budget of the earth system it proves useful to divide, Q_0, by the ratio of Earth's surface area to its projected area on the solar sky. This ratio, 4.0034, deviates only slightly from that of a perfect sphere, and along with an assumed amplitude of the solar cycle of $1.5\,\mathrm{W\,m^{-2}}$ is used to define the effective total solar irradiance, $Q_\star = 340.1 \pm 0.1, \mathrm{W\,m^{-2}}$ that is, on average, incident on the top of the atmosphere. For most purposes it is sufficiently accurate to take $Q_\star = 340\,\mathrm{W\,m^{-2}}$.

In stationarity, the solar irradiance at the top of the atmosphere is balanced by a lost of enthalpy through the emission of longwave radiation, E. As mentioned above, Earth's temperature dictates that this energy is predominantly carried at longer wavelengths, and it peaks in the infrared with wavelengths near $10\,\mu$m. Adopting a sign convention so that an inward directed source of energy is taken to be positive, the heat (or enthalpy) budget of the Earth system can be written as,

$$\dot{H} = Q_\star \gamma - E, \tag{1}$$

where γ is the co-albedo, i.e., $\gamma = 1 - \alpha$ where α is the ratio of the incoming to the outgoing solar irradiance averaged over all sun angles, i.e., the spherical albedo, and H is the total enthalpy (heat content) of the Earth system. In stationarity (balance), H is constant, so that $\dot{H} = 0$, where the *dot* notation is used to denote time differentiation. Given a heat capacity for the system as a whole, usually taken to be the heat capacity of the atmosphere, ocean and upper portions of the land surface, H can be related to the temperature of the earth and the distribution of matter (water) in different phases.

Clouds play an important role in the energy budget because the modify the TOA irradiances. To make their role explicit it proves convenient to decompose Q, E and γ into terms associated with cloudy, versus clear skies, for example by expressing the albedo as a combination of the clear and cloudy sky albedo,

$$\alpha = \alpha_0(1 - A_\mathrm{c}) - \alpha_\mathrm{c} A_\mathrm{c}, \tag{2}$$

the solar irradiance becomes

$$Q = Q_\star \left[1 - \alpha_0(1 - A_\mathrm{c}) - \alpha_\mathrm{c} A_\mathrm{c}\right]. \tag{3}$$

Likewise, because clouds are so opaque to long wave radiation, E depends strongly on the presence of clouds, particularly so for the case of clouds whose temperature is very different than that of the underlying surface. This motivates the decomposition of E into a part attributable to clouds and a part that is what one would expect in their absence, so that

$$E = E_0(1 - A_\mathrm{c}) + E_\mathrm{c} A_\mathrm{c}, \tag{4}$$

where in general one expects both E_0 and E_c to depend on the temperature of the emission source, or a weighed temperature of a source region.

It is important to realize that, as introduced above, parameters such as A_c and α_c are effective or bulk quantities. For instance α_c is not the average cloud albedo, but rather the albedo a cloud would require to reflect the increment in the net radiation associated with cloudy skies. To illustrate this concept, imagine an earth in which the solar illumination is uniform over the surface, but whose surface albedo varies in away that correlates with cloudiness. The effect of these correlations, and how one defines the clear sky albedo, must be accounted for in the definition of the effective cloud fraction. If the incident radiation also correlates with cloudiness this also must be accounted for. For instance, if it were only cloud at night, then for purposes of calculating, Q, the effective cloud amount must be zero.

2 Earth's Energy Flows

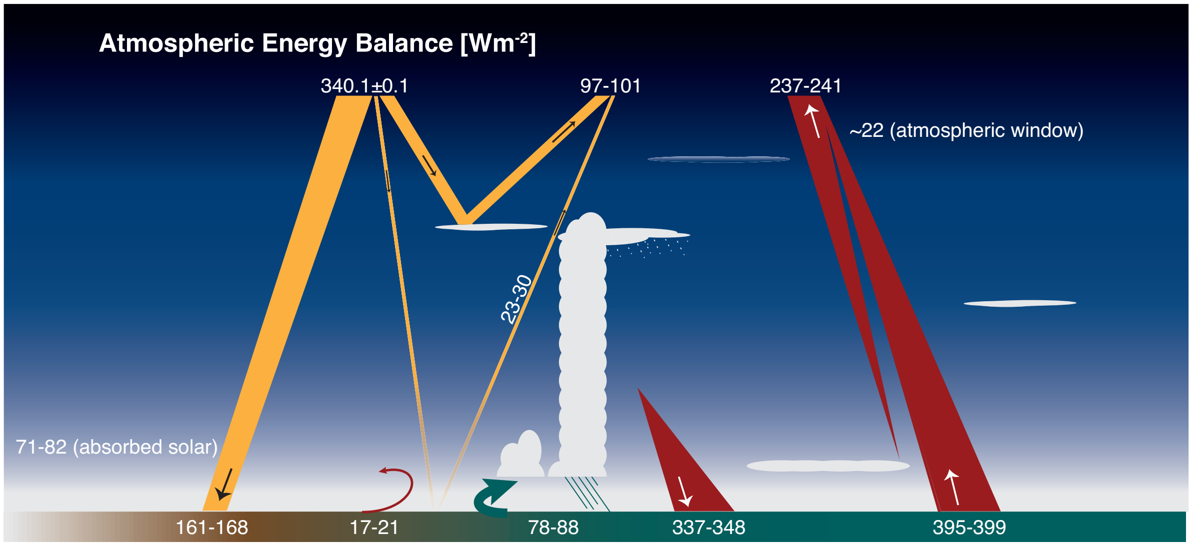

Figure 1: Earth's global and annual mean top-of-the-atmosphere (TOA) and surface energy budget. Values are presented as a two-sigma range (which is to say the authors assign a roughly 68 % likelihood that the actual values fall within the stated range). These values are subjectively determined based on a review of the literature, and complemented by global simulations, as derived by Stevens and Schwartz 2012.

Measurements have been indispensable to the advancing understanding of energy flows through the climate system. Current understanding of these flows is summarized in Fig. 1, which we have constructed based on the available literature (especially Trenberth et al., 2009; Kato et al., 2012; Stephens et al., 2012). The energy flows are more certain at the top of the atmosphere than at the surface, as measurements at the top of the atmosphere have benefitted greatly from advances in satellite remote sensing. Measurements from the Earth Radiation Budget (ERB) program (Ramanathan, 1987) and now those from the Clouds and Earth's Radiant Energy System (CERES) mission (Wielicki et al., 1996) have convincingly shown that Earth reflects less shortwave radiation and emits more longwave radiation than previously thought (cf., Table 1) and have contributed to closure of the top-of-atmosphere energy budget to within a few watts per square meter.

Also contributing to the reduction of uncertainty in the balance of energy flows at the top-of-the-atmosphere are new measurements of total solar irradiance and ocean heat uptake. Because the atmosphere has a relatively small heat capacity, an imbalance of energy flows at the TOA can be sustained only an increase in ocean enthalpy, augmented to lesser extent by melting of the cryosphere and warming of the land surface and the atmosphere. The rate of heating of the top 700 m of the world ocean over the last sixteen years, as inferred from temperature measurements and expressed per the area of the entire planet, is $0.64 \pm 0.11\,\mathrm{W\,m^{-2}}$ (90 % confidence interval). Sparser measurements extending to ocean depths of 3 km reported by Levitus et al. (2005) suggest that the upper ocean takes up about three-quarters of the ocean heating. Levitus et al. (2005) also estimate that the contributions of other heat sinks, including the atmosphere, the land, and the melting of ice, can account for an additional $0.04\,\mathrm{W\,m^{-2}}$; Based on these estimates the flux imbalance at the TOA is estimated to be $0.9 \pm 0.3\,\mathrm{W\,m^{-2}}$.The uncertainty is based on the 90 % confidence interval given by Lyman and the assumption that the relative uncertainty in the deep ocean heat-

Source	H-1954	L-1957	R-1987	K-1997	Z-2005	T-2009	S-2012	Fig. 1
Top of Atmosphere								
Incident SW	338	349	343	342	342	341	340	340
Reflected SW	115	122	106	107	106	102	98	100
Outgoing LW	223	227	237	235	233	239	239	239
Surface								
Absorbed SW	160	167	169	168	165	161	168	162
Downward LW	352	334	327	324	345	333	347	342
Upward LW	398	395	390	390	396	396	398	397
Latent heat flux	82	68	90	78	n/a	80	88	86
Sensible heat flux	30	33	16	24	n/a	17	24	20

Table 1: Estimates of Earth's energy budget, subjectively determined based on a review of the existing literature and the best estimates of the net imbalance at the surface and top-of-atmosphere. References to prior estimates are: Houghton (2011); London (1957); Ramanathan (1987); Kiehl and Trenberth (1997); Trenberth et al. (2009); Stephens et al. (2012). Because both the H-1954 and L-1957 estimates were for the northern hemisphere only, these estimates have been rescaled using the ratio of the global versus the northern hemisphere average from a high resolution AMIP simulation using the ECHAM6 model. For basis of present estimates see Appendix A.

uptake estimates and in the estimates of heating by other components of the Earth system are about 50 %.[1] For reference, in constructing the energy-balanced version of the CERES data Loeb et al. (2009) estimated the surface heat uptake to be $0.85\,\mathrm{W\,m^{-2}}$ similar to the 0.9 Wmm estimated here and employed by Trenberth et al. (2009) based on a somewhat different line of reasoning. In summary, although there is a rather larger, $6.5\,\mathrm{W\,m^{-2}}$, inherent uncertainty in measurements of the reflected shortwave and emitted longwave radiation at the TOA, associated principally with uncertainty in the absolute calibration of the CERES instruments (Loeb et al., 2009), the net irradiance at the TOA is constrained by improved measurements of increases in ocean enthalpy.

The surface energy budget is distributed over several terms, each of which exhibits uncertainty that is several-fold greater than the uncertainty in the net budget at TOA. As pointed out by Trenberth et al. (2009), if each of the terms in the surface energy budget is estimated individually, in isolation of the others, an imbalance can arise in the net surface flux that is as much as $20\,\mathrm{W\,m^{-2}}$; this is more than an order of magnitude greater than current measurement-based estimates of the rate of heat uptake by the ocean, cryosphere and land. In absolute terms the uncertainty is largest for the longwave irradiance downwelling at the surface and for the latent heat flux (precipitation). Here too, new measurements and improved modelling are beginning to improve understanding. Active remote sensors such as the cloud profiling radar flown as part of the CloudSat mission (Stephens et al., 2008) and the Cloud-Aerosol Lidar with Orthogonal Polarization flown as part of the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation mission (Winker et al., 2010) are providing improved estimates of the vertical distribution of clouds, particularly cloud base. These measurements are crucial for estimates of the downward longwave irradiance, which are based on radiative transfer modeling given a realistic atmospheric state. Using these measurements Kato et al. (2012) estimate a downward longwave irradiance at the surface of $345.4 \pm 6.9\,\mathrm{W\,m^{-2}}$ which is considerably larger than the value derived by Trenberth et al. (2009) from the residual of the surface energy balance given existing precipitation climatologies. CloudSat and surface-based measurements also suggest that existing precipitation climatologies underestimate light rain from warm clouds, which is common over the tropical

[1] Other sources of energy – geothermal, combustion of fossil fuel and nuclear production – are yet an order of magnitude smaller.

ocean (Nuijens et al., 2009; Stephens et al., 2012). Overall the uncertainty in the surface energy budget, as presented in Fig. 1 is somewhat smaller than that presented by Stephens et al. (2012), as global modelling (including reanalysis) estimates constrained by observations is given somewhat more weight, particularly for the upward turbulent fluxes and the long-wave irradiance.

3 Clouds

4 Precipitation

5 Cloud Radiative Effects

A framework for considering cloud feedbacks in the climate system, first formulated by Schneider (1972), introduced the idea of cloud forcing (see also Cess, 1976; Ramanathan, 1987). Identifying the subset of short- and longwave radiative fluxes associated with cloud-free scenes, and assuming that three-dimensional effects associated with neighboring clouds can be neglected, permits determining "cloud radiative forcing" as the difference between the all-sky or actual radiative flux and the contribution from scenes having skies identified as cloud free. Here, as before, the term "cloud-forcing" is used to denote a change in radiative flux due to a change in atmospheric composition, namely clouds. However, because this terminology does not allow one to distinguish the radiative effect of the totality of clouds, from the radiative perturbation that would accompany a perturbation in cloud properties, it proves useful to distinguish between the cloud-radiative effect, CRE, as the radiative effect of the background state of cloudiness, and reserve the phrase "radiative forcing" for radiative perturbations driven by externally imposed changes.

The definition of the cloud radiative effect can be made precise as follows. Here we return to Eqs. (3) and (4)

$$\begin{aligned} Q &= Q_\star\left[1-\alpha_0(1-A_\mathrm{c})-\alpha_\mathrm{c}A_\mathrm{c}\right] && (5)\\ E &= E_0(1-A_\mathrm{c})+E_\mathrm{c}A_\mathrm{c} && (6)\end{aligned}$$

The short- and longwave components of the CRE, denoted by superscripts (Q) and (E) respectively, follow naturally as the difference between the all-sky radiative flux and the fluxes which would be manifest in the absence of clouds, i.e.,

$$\begin{aligned} F_\mathrm{c}^{(Q)} &= -Q_\star(\alpha_\mathrm{c}-\alpha_0)A_\mathrm{c} && (7)\\ F_\mathrm{c}^{(E)} &= (E_\mathrm{c}-E_0)A_\mathrm{c} && (8)\end{aligned}$$

Given our sign convention, and because α_c is generally greater than α_0 whereas E_c is generally less in magnitude than E_0 (both being negative), the shortwave CRE is negative and the longwave CRE is positive. Both quantities increase in magnitude with cloud amount, A_c. The net CRE, F_c, is given by $F_\mathrm{c}^{(Q)}+F_\mathrm{c}^{(E)}$. A secular increase in the magnitude of $F_\mathrm{c}^{(Q)}$ would exert a cooling influence on the Earth system, whereas an increase in the magnitude of $F_\mathrm{c}^{(E)}$ would exert a warming influence. It should be emphasized that the CRE depends not just on the properties of the cloudy fraction of the planet but on the *differences* between the cloudy and cloud-free portions of the planet. As reviewed by Loeb et al. (2009) the application of this concept to various data sets shows the shortwave CRE to range from -45.4 to -53.3 $\mathrm{W\,m^{-2}}$ and the longwave CRE to range from 26.5 to 30.6 $\mathrm{W\,m^{-2}}$; the associated net CRE from these prior estimates ranges from -16.7 to -24.5 $\mathrm{W\,m^{-2}}$. The CERES EBAF data, upon which the TOA estimates in Fig. 1 are largely based, give a shortwave cloud radiative effect of -47.1 $\mathrm{W\,m^{-2}}$ and a longwave CRE of 26.5 $\mathrm{W\,m^{-2}}$; overall clouds, more precisely cloudy-scenes, exert a net cooling influence on the Earth system of about 20 $\mathrm{W\,m^{-2}}$.

A subtlety of the cloud forcing concept is that the CRE, as defined above, depends not only on the properties of the cloud-free scenes, but also on other radiation-influencing constituents of the climate system that might be correlated with the presence of clouds. Hence it is really a cloudy-scene radiative effect, i.e., the difference between the total radiative flux, and that which would obtain if cloud-free conditions always prevailed. The emphasis on cloudy-scenes rather than clouds is because the former admits the possibility that the atmosphere in cloudy scenes is systematically different from the atmosphere in cloud-free scenes. For instance, if the atmosphere in cloudy scenes tends to be more humid than in cloud-free scenes then this difference in humidity will, through the definition of cloud radiative effect (8), be interpreted as an effect of clouds. This distinction has important consequences for how cloud radiative effects are calculated in models versus in measurements. In models the radiative transfer calculation is usually performed twice, the second time with the clouds removed from the input. The first call to the radiation defines the all-sky radiative flux, the second call defines the clear-sky radiative flux and their difference is called the cloud radiative effect. In observations one cannot usually remove clouds[2], so the clear-sky radiative flux is estimated based on scenes where no clouds are identified in the first place, rather than based on all scenes but with removal only of the clouds from those scenes in which they occur. If the atmosphere around the clouds is different (for instance more humid) in the cloudy scenes, as one might expect, this leads to differences between the two estimates of CRE. Such effects were shown by Sohn et al. (2010) to lead to systematic discrepancies between the observed longwave CRE and that calculated by models. Such discrepancies can, however, be readily overcome by calculating the CRE in models in the same way as is done in observations, by compositing over cloud free scenes when determining the clear-sky flux.

[2]It might appear that there are exceptions. Intrieri et al. (2002) and Mauritsen et al. (2011) calculate the clear-sky radiative fluxes using measurements of the atmosphere in cloudy conditions, but do not include the clouds in the radiative transfer calculations. However this is not an actual measurement of radiative effects, but rather a calculation of radiative effects given measured atmospheric properties.

6 Forcing, Response, Adjustment and Feedbacks

In addition to providing understanding of energy flows in the current climate, consideration of the energy budget also provides a framework for understanding climate change. This framework, which has developed over the last thirty years, rests on the assumption that changes in the globally averaged surface temperature can be xlinearly related to a radiative forcing, F. The constant of proportionality between the forcing and the response is called the equilibrium climate sensitivity, $S_{\rm eq}$, which can be formally defined as the steady-state change in $T_{\rm s}$ (the globally averaged near-surface air temperature) that would result from a sustained change in a radiative flux component of the Earth energy budget at the TOA (forcing), normalized to that flux change, with unit: $\mathrm{K}\,(\mathrm{W\,m^{-2}})^{-1}$. That is, the equilibrium sensitivity is the proportionality constant between the steady-state change in surface temperature and the applied forcing,

$$\delta T_{\rm s} = S_{\rm eq} F. \tag{9}$$

The symbol δ is used to imply a small change, presumably because the forcing F is also small. Because it codifies the sensitivity of such an essential feature of the climate state, the globally averaged surface temperature, Eq. 9 provides a powerful description of past and prospective future climate change. To the extent that other properties of the climate system scale with its value, $S_{\rm eq}$ assumes an even broader significance. For these reasons, determination of $S_{\rm eq}$ has evolved into a central focus of climate science (e.g., Knutti and Hegerl, 2008).

6.1 Forcing

An external forcing, F can be introduced as an additional term in Eq. 9, so that

$$\dot{H} = Q_\star \gamma - E + F, \tag{10}$$

where the sign of F is chosen so that a positive value implies an increase in the system enthalpy. In so doing we note that F is considered to be an increment on the background, or normal, forcing Q of the earth system, in essence F defines a forcing anomaly. For instance, as a result of a change in $Q_\star$ due to a change in the sun itself, or through the reduction (for instance from space borne mirrors) in the amount of solar irradiance that reaches the top of the atmosphere

The idea of the forcing arising as an anomaly in the energy budget of the system can be generalized to include other sources. For instance, volcanism, by introducing sulfur in the stratosphere which leads to the production of stratospheric aerosol, acts much like space-borne mirrors, to reduce the co-albedo, γ. Likewise the long wave irradiance, E, at the top of the atmosphere can be perturbed by changing the composition of greenhouse gases. Instantaneously doubling the amount of CO_2 in the atmosphere is expected to change the longwave irradiance at TOA by about $3\,\mathrm{W\,m^{-2}}$. This idea of equating an internal perturbation to the system, and its associated radiative perturbation, with a forcing can, however, be quite confusing. Consider the case wherein a compositional change leads to a decrease in the emission of longwave radiation, and let $F = \delta E$. With the passage of time one expects the system to adjust its temperature and return to stationarity, and assuming the albedo does not change, in this new stationary state $Q_\star\gamma$ will remain the same, but E must increase by an amount δE so that $E + F = Q_\star\gamma$ which of course would return the perturbed system to stationarity. Hence this idea of forcing has intrinsic to it the idea of a natural state, i.e., that state in which $F = 0$.

These ideas can be developed more precisely. Let χ denote the concentration of greenhouse gases, and let E depend on χ and on (among other quantities) the globally averaged surface temperature T_s. Then a compositional change in the atmosphere can be denoted by a perturbation to χ, or $\delta\chi$, and this leads to a forcing, as

$$F = -\delta E = -\frac{\partial E}{\partial \chi}\delta\chi, \tag{11}$$

so that Eq. 10 becomes

$$\dot{H} = Q_{\star}\gamma - E + \frac{\partial E}{\partial \chi}\delta\chi. \tag{12}$$

As H increases, T_{s} increases, leading to a change a further change in E, and eventually a new equilibrium is reached when

$$\frac{\partial E}{\partial T_{\mathrm{s}}}\delta T_{\mathrm{s}} = \overbrace{\frac{\partial E}{\partial \chi}\delta\chi}^{F}. \tag{13}$$

By comparing Eq. 13 with Eq. (9) it becomes apparent that, in this case, the climate sensitivity can be directly equated with the partial derivative of the emission with respect to the surface temperature,

$$S_{\mathrm{eq}} = \left(\frac{\partial E}{\partial T_{\mathrm{s}}}\right)^{-1}. \tag{14}$$

This raises the possibility that S_{eq} depends on the surface temperature, as the partial derivative must be evaluated at this temperature.

Although S_{eq} could depend on the working temperature, T_{s} the generality of the concept of a climate sensitivity rests on it being independent of the forcing. This is not generally the case, and care must be exercised when discussing the forcing and the response that arise from different types of perturbations. The issue is further complicated by the dependence of the response of the system to a given forcing on how one defines the system. A system in which land-ice, the biosphere, and the ocean circulation are allowed to change with changing surface temperature will respond differently than one in which these slowly responding components are held fixed. In an effort to address these potential ambiguities, some authors speak of the Charney sensitivity as the value of Seq that would arise from a doubling of CO_2 in the absence of changes to the land biosphere, cryosphere or land surface. However, from a historical perspective, the climate sensitivity, S_{eq}, is synonymous with the Charney sensitivity.

6.2 Adjustment

The concept of adjustment has been introduced to help overcome the lack of universality in a radiative forcing defined by the instantaneous change in the TOA irradiance. Initially the idea of adjustment was introduced to account for rapid changes in the stratosphere that occur when the CO_2 concentration is perturbed. As CO_2 increases the stratosphere cools, relatively rapidly, thereby further reducing the long wave irradiance at the TOA, and making the effective forcing from a doubling of CO_2 somewhat stronger than it would be if these adjustments were not considered. By allowing for these adjustments when calculating the forcing, one finds that that there is a better correspondence in the response of the system to $1\,\mathrm{W\,m^{-2}}$ of solar forcing and $1\,\mathrm{W\,m^{-2}}$ of adjusted CO_2 forcing than there is to $1\,\mathrm{W\,m^{-2}}$ of instantaneous CO_2 forcing.

Other adjustments also have been identified. For instance cloudiness adjusts rapidly in response to a change in CO_2. These adjustments occur because the long wave irradiance, which depends on the amount and distribution of greenhouse gases, directly effects some cloud systems. Stratiform cloud layers are directly effected, but other cloud forms may also change because the thermal stratification in the atmosphere depends on the distribution of irradiance in the troposphere, and this in turn effects cloudiness. Because these and other adjustments are perturbation specific, and relatively quick, it proves useful to associated F in Eq. (9) with the adjusted forcing. However, because F now depends on a more complex suite of processes than it would if it were to be defined based on the instantaneous radiative response to a change in atmosphere composition, the translation of a compositional change of the atmosphere into a radiative forcing becomes non-trivial and much more uncertain, with the rapid adjustment of clouds being an important source of uncertainty.

Mathematically these ideas can be formalized by considering a slightly more expansive example than what that provided earlier. If we let E depend on a hidden variable, say the atmospheric temperature lapse rate Γ. Then if Γ adjusts, on relatively short timescales, to compositional perturbations, it is straightforward to show that the forcing must include an additional contribution from this adjustment, namely,

$$F = \left(\frac{\partial E}{\partial \chi} + \frac{\partial E}{\partial \Gamma}\frac{\partial \Gamma}{\partial \chi}\right)\delta\chi \tag{15}$$

It turns out that by defining the forcing in this way the system responds much more universally to a given forcing, thus allowing one to compare the climatic impact of a change in greenhouse gases, to a change in the solar constant. The term related to the static stability in Eq. (15) has been given different names in the literature, but is best seen as an adjustment. It is distinct from the response in so far as changes in the static stability are assumed to be independent of changes in the globally averaged surface temperature, T_s.

6.3 Feedback

A feedback can be thought of as resulting from a dependence of our effective description of the climate system on the surface temperature. For instance, suppose that Γ depended on both T_s and χ. Then it is straightforward to show that

$$\overbrace{\left(\frac{\partial E}{\partial T_\mathrm{s}} + \frac{\partial E}{\partial \Gamma}\frac{\partial \Gamma}{\partial T_\mathrm{s}}\right)}^{S_\mathrm{eq}^{-1}}\delta T_s = \overbrace{\left(\frac{\partial E}{\partial \chi} + \frac{\partial E}{\partial \Gamma}\frac{\partial \Gamma}{\partial \chi}\right)}^{F} \tag{16}$$

In this example we can speak of a lapse-rate feedback, as the change in the lapse rate with temperature modifies the overall response of the system. likewise, if γ also depends on temperature then the perturbation analysis of Eq. 10 results in

$$\overbrace{\left(\frac{\partial E}{\partial T_\mathrm{s}} + \frac{\partial E}{\partial \Gamma}\frac{\partial \Gamma}{\partial T_\mathrm{s}} + Q_\star\frac{\partial \gamma}{\partial T_\mathrm{s}}\right)}^{S_\mathrm{eq}^{-1}}\delta T_s = \overbrace{\left(\frac{\partial E}{\partial \chi} + \frac{\partial E}{\partial \Gamma}\frac{\partial \Gamma}{\partial \chi}\right)}^{F}. \tag{17}$$

Likewise, if E is assumed to additionally depend on the absolute humidity, q, and q is assumed to vary with temperature, as would be the case if relative humidity were to remain constant, the inverse climate sensitivity would take the form

$$S_\mathrm{eq}^{-1} = \left(\frac{\partial E}{\partial T_\mathrm{s}} + \frac{\partial E}{\partial \Gamma}\frac{\partial \Gamma}{\partial T_\mathrm{s}} + \frac{\partial E}{\partial q}\frac{\partial q}{\partial T_\mathrm{s}} - Q_\star\frac{\partial \gamma}{\partial T_\mathrm{s}}\right). \tag{18}$$

The climate sensitivity can be expressed more intuitively if the expression for the emission is given a form that makes the temperature dependence more explicit. In analogy to black body radiation we write

$$E = \varepsilon\sigma T_s^4, \quad \text{where} \quad \varepsilon = \varepsilon(q, T_\mathrm{s}). \tag{19}$$

Thus

$$\frac{\partial E}{\partial T_\mathrm{s}} = 4\varepsilon\sigma T_s^3 = 4\frac{Q_\star\gamma}{T_\mathrm{s}}, \tag{20}$$

where the second equality follows from the partial derivative being evaluated about the non-perturbed stationary state, in which energy balance dictates that $E = Q_\star\gamma$. Substituting the above into the expression (18) allows us to write the inverse climate sensitivity as

$$S_\mathrm{eq} = \frac{T_\mathrm{s}}{4Q_\star\gamma}\left[1 + \frac{T_\mathrm{s}}{4\varepsilon}\frac{\partial\varepsilon}{\partial\Gamma}\frac{\partial\Gamma}{\partial T_\mathrm{s}} + \frac{T_\mathrm{s}}{4\varepsilon}\frac{\partial\varepsilon}{\partial q}\frac{\partial q}{\partial T_\mathrm{s}} - \frac{T_s}{4\gamma}\frac{\partial\gamma}{\partial T_\mathrm{s}}\right]^{-1}. \tag{21}$$

The terms in the brackets can be thought of as the gain, that arise from the feedbacks in the system. That is we can write the climate sensitivity as a product of the climate sensitivity that would arise in the absence of feedbacks and a gain, G,

$$S_\mathrm{eq} = S_\mathrm{eq}|_{f=0}\, G \quad \text{with} \quad G \equiv \frac{1}{1-f} \tag{22}$$

where the term f appearing in the definition of the G defines the feedback strength. In our example,

$$f = \frac{T_s}{4\gamma}\frac{\partial\gamma}{\partial T_\mathrm{s}} - \frac{T_\mathrm{s}}{4\varepsilon}\frac{\partial\varepsilon}{\partial\Gamma}\frac{\partial\Gamma}{\partial T_\mathrm{s}} - \frac{T_\mathrm{s}}{4\varepsilon}\frac{\partial\varepsilon}{\partial q}\frac{\partial q}{\partial T_\mathrm{s}}. \tag{23}$$

So the total feedback is the sum of individual feedbacks arising form the temperature dependence of the emissivity via changes in water vapor and the temperature lapse rate, and the temperature dependence of the co-albedo. The feedback strength is dimensionless, and the system is said to be stable if the net feedback strength is less than unity. In this example the feedbacks we identify are the water-vapor feedback, the lapse rate feedback, and albedo feedbacks.

6.4 Meta-points on forcings and feedbacks

The above analysis is not fundamental. Rather it *interprets* the behavior of a complex system through the use of a simple model, namely Eq. (10) accompanied by auxiliary hypotheses (for example Eq. 19) on the independent variables that control the irradiances. The interoperation one makes can, and often does, depend on the model which is being used to interpret the behavior of the system. For instance, the water vapor feedback arises because it is assumed that humidity distribution effects the emission, and this in turn depends on temperature. Obviously there are many other possible contributions to the response of the system that are not identified in Eq. 23, for instance the horizontal temperature distribution might impact the net emission, and if this changes with changing T_s it, independent of changes in surface albedo, would constitute a feedback. Because this feedback is not explicitly identified in Eq. 23 its contribution to the total response must be carried by other terms. Additionally, some terms that are independent, might not be so. For instance in the analysis leading to Eq. (23) it assumes that the lapse-rate, Γ and the humidity q are independent variables. But they need not be, and both may change as a result of another process which in turn is responsible for the underlying sensitivity to temperature. Failing to account for this co-dependence can result in differences among models being mis-interpreted. Finally, the relationships among macroscopic variables are statistical. So to determine how the emissivity changes with absolute humidity one might conduct two simulations, one in which the humidity is held constant in the calculation of the radiative fluxes, and the other in which the h humidity is allowed to vary. But such an approach is easier said than done, and the answer one arrives at can depend on your choice of coordinate system. A humidity that is prescribed to be constant in a coordinate system where height is the physical coordinate will behave differently than a humidity that is prescribed to be constant in a coordinate system where pressure, or potential temperature, is taken as the physical coordinate. Despite these ambiguities, feedback analysis can be useful to get an idea as to how different processes contribute to the response of the system, the results of such an analysis just require care in their interpretation.

The main difference between adjustments and feedbacks, are that feedbacks are adjustments that are mediated through (correlated with) surface temperature changes. The term “feedback” is often used loosely in the climate change discourse to refer to a contribution of a process to the response of the system. So doing misses the essential point of a feedback, in that it describes how the response of a system to a given forcing changes that forcing. If the response of the system is determined to be globally averaged surface temperature, then feedback describe those changes in the system that arise because the globally averaged surface temperature is changing and which in turn modify the TOA irradiances. In contrast, an adjustment is a change in the system that occurs independently of changes in surface temperature. On a practical level adjustments and feedbacks are distinguished through a timescale argument, as adjustments are assumed to happen rapidly, and because feedbacks depend on globally averaged surface temperate, which given the heat capacity of the system evolve slowly.

7 Cloud-aerosol interactions

8 Modeling clouds in the climate system

9 Modeling convection in the climate system

10 Clouds and climate change

11 Precipitation and climate change

12 Landsurface interactions

13 Teleconnections

References

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A Taylor Series and perturbation analysis

Let $f(x)$ be a function of x, and assume that it can be represented by a convergent power series (which for most physical functions of interest is a good assumption) such that

$$f(x) = \sum_{n=0}^{\infty} c_n (x-a)^n. \tag{24}$$

The constants c_n define the function. If the series is convergent in some interval about a, then successive terms in the series have diminishing magnitude (by the alternating series theorem) and it also follows that the constants, c_n, can be written in terms of derivatives of the function, namely

$$c_n = \frac{f^{(n)}(a)}{n!}, \tag{25}$$

where $f^{(n)}$ denotes the nth derivative of f. Note that $c_0 = f(a)$. When the constants, c_n, are defined in this way Eq. (24) is called the Taylor series for f about the point a.

Based on the Taylor series one can ask how the function, f changes for a small change in the independent variable x. From the Taylor series approximation it follows by substitution that

$$\delta f = f(a+\delta x) - f(a) = \sum_{n=1}^{\infty} \frac{f^{(n)}(a)}{n!} (\delta x)^n. \tag{26}$$

In the interval over which the series approximation converges to the function the leading order term is the first term, and this term increasingly dominates as δx goes to zero. For sufficiently small perturbations we can thus write,

$$\delta f \approx \left.\frac{\mathrm{d}f}{\mathrm{d}x}\right|_a \delta x \quad \text{(for all } a\text{)}. \tag{27}$$

The vertical bar with the subscript a denotes the evaluation of the derivative at the point a. This idea generalizes to more than one dimensions, so that $g(x,y)$ can be approximated, or linearized about an arbitrary point (a,b) as

$$\delta g \approx \left.\frac{\partial g}{\partial x}\right|_a \delta x + \left.\frac{\partial g}{\partial y}\right|_b \delta y. \tag{28}$$

As an example, consider black body radiation, where by the emission of thermal radiation is a function of temperature, T so that $E(T) = \sigma T^4$, with $\sigma = 5.67 \times 10^{-8}\,\mathrm{W\,m^{-2}\,K^{-4}}$. If we know the emission, E_0, at some base temperature T_0, for a small change in temperature we can estimate how the emission will change as

$$\delta E = 4\sigma T_0^3 (T - T_0), \quad \text{equivalently} \quad \frac{\delta E}{E_0} = 4\frac{\delta T}{T_0}. \tag{29}$$

This type of analysis is used frequently to understand how a function will change due to a small perturbation of its arguments. Because many functions that we deal with take the form of a power-law, in their respective terms, such that

$$g(x,y) = \gamma x^{\alpha} y^{-\beta}, \tag{30}$$

it is useful to note that these linearize about some arbitrary point (x,y) as

$$\frac{\delta g}{g} \approx \alpha \frac{\delta x}{x} - \beta \frac{\delta y}{y}. \tag{31}$$

B Cloud Slam Papers

The following is intended to be a brief annotated bibliography of some of the more important papers related to a climate system view of clouds and convection. It is an evolving list, contributions and suggests are welcome.

- Forcing, adjustment and feedbacks:Forster and Gregory (2006); Gregory and Webb (2008) and the introduction to feedback analysis by Schwartz (2011).
- Cloud radiative effects by Ramanathan (1987) and the differences between observations and models as discussed by Sohn et al. (2010) as well as the classic study of cloud susceptibility to perturbations by Hartmann and Short (1980)
- For cloud parameterization, simple statistical relations are discussed by Slingo (1987); Klein and Hartmann (1993), modeling approaches by Sundqvist (1978); Tompkins (2009)
- Surface energy budget from measurements Kato et al. (2012) and constraints resulting from properties of radiative transfer Stephens and Hu (2010), and a forcing-feedback view by Andrews et al. (2009), also a view of GCM behavior in Richter and Xie (2008)
- Changes in the hydrological cycle: Held and Soden (2006)
- Biases in the large-scale representation of precipitation, e.g., the double ITCZ problem Lin (2007).
- Cloud feedbacks: Somerville and Remer (1984); Bony and Dufresne (2006), Hartmann and Larson (2002), Brient and Bony (2011); Clement et al. (2009); Rieck et al. (2012); Zelinka and Hartmann (2010, 2011); Webb et al. (2012)
- Cloud adjustment to CO_2 Wyant et al. (2012) and the aerosol Twomey (1974); Albrecht (1989); Pincus and Baker (1994); Stevens and Feingold (2009); Lohmann and Feichter (2005).
- Land surface interactions associated with precipitation soil-moisture feedbacks Hohenegger et al. (2009)