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On the stability of the Earth's radiative energy balance: Response to the Mt. Pinatubo eruption

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Received 4 July 2006; revised 27 September 2006; accepted 7 November 2006; published 15 December 2006.

[1] A volcanic eruption provides a natural experiment in which time constants for the onset and decay of the consequent radiative perturbation may be measured. The radiative and atmospheric responses also provide insight into how the top of atmosphere net balance of energy responds to perturbations. We analyse the response of the atmospheric humidity, temperature and radiative fields to the eruption of Mt Pinatubo to determine time constants for the growth and the decay of perturbations in these fields. We are able to distinguish between processes that respond directly to the insertion of aerosols into the atmosphere, and those, such as changes in the humidity and temperature, that involve slower dynamical processes and therefore have longer response times. The physical basis for these observations is discussed, and it is suggested that a valuable test of coupled climate models should be that they reproduce these response times, and the associated flux anomalies. **Citation:** Harries, J. E., and J. M. Futyan (2006), On the stability of the Earth's radiative energy balance: Response to the Mt. Pinatubo eruption, *Geophys. Res. Lett.*, *33*, L23814, doi:10.1029/2006GL027457.

1. Introduction

[2] The balance of radiative energy at the top of the atmosphere (TOA), measured by the net flux, F_N, (the difference between down-welling shortwave (SW) radiation absorbed by the Earth, and the up-welling longwave (LW) radiation emitted back to space) is a question of current interest. Although the Earth is a long way from thermodynamic equilibrium [*Kleidon and Lorentz*, 2005], as evidenced by the large entropy production due to the many processes which ultimately transform the incoming short wave energy into the outgoing long wave energy stream, it is of interest to understand how and why the TOA net flux might drift in and out of balance, and if does so, on what time scales this occurs. Satellite measurements of this balance [*Wielicki et al.*, 2002] have been used to indicate decadal-scale variations in the magnitude of the TOA net flux of up to about 5 W m^{-2} in the tropical zone (20N–20S), and model simulations [*Hansen et al.*, 2005] have indicated that, due to the delay in temperature response to global warming, caused presumably by the oceans, a global imbalance of about +1 W m^{-2} in the net flux has developed between about 1960 and 2000. However, it is not yet clear whether this is apparent in the satellite record, possibly due to such small changes being undetectable given current measurement capability. It is an interesting question to ask what evidence there is for a steady state of the climate system, though this is a complicated issue somewhat beyond the scope of this paper (to be pursued elsewhere). Rather, we can address here questions such as 'how far out of balance does F_N drift, and how fast?' We need more (and more accurate) observations and model simulations to understand this problem. If the climate system regains TOA balance quickly after a perturbation, perhaps the observation of the global F_N is not a useful way to study global climate change (although almost certainly of value on a more regional scale). A volcanic eruption provides a powerful way of exploring these questions, and this is the purpose of this paper.

[3] The direct effect of a change in radiative forcing, for example due to increasing greenhouse gases such as CO_2, CH_4 and others, or due to atmospheric aerosols, is now well established theoretically (see the extensive work reported in the Intergovernmental Panel on Climate Change (IPCC) reports [e.g., *Intergovernmental Panel on Climate Change*, 2001]), and this theory has been given direct confirmation recently from observations of the spectrum of thermal emitted radiation from the Earth, which indicates that the greenhouse forcing of the Earth has, indeed, been changing over the past 30 years, in accordance with known changes of the greenhouse gas composition of the atmosphere [*Harries et al.*, 2001]. However, the challenge remains to understand the actual rate and magnitude of climate change, in response to a change in forcing, which, of course, depends strongly on feedback processes, for example the water vapour [*Kiehl and Trenberth*, 1997; *Held and Soden*, 2000; *Harries*, 2000], and the cloud feedbacks [*Cess et al.*, 1990; *Hansen et al.*, 2002]. There is considerable uncertainty associated with the description of these feedback processes as represented in models, especially those processes associated with clouds [e.g., *Webb et al.*, 2001; *Hansen et al.*, 2002]: this, in turn, gives rise to uncertainty about the accuracy of climate change predictions.

[4] The speed with which the climate system can respond to a perturbation through these various feedback processes depends on a number of factors. These include [*Houghton*, 2002] the thermal mass of the component of the climate system concerned (small for the upper atmosphere, very large for the deep oceans), and whether heat energy can be transported directly by motions (as in the atmosphere and the oceans) or only by conduction (as in the land). We can expect the response of the system as a whole to be a complex mix of the responses to all processes. Efforts to measure and model the magnitude and duration of perturbations to the TOA net flux indicate that the net flux is

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0094-8276/06/2006GL027457$05.00

restored quickly to zero, and does not depart from zero by more than a few W m^{-2} [*Wielicki et al.*, 2002; *Hansen et al.*, 2005].

2. Responses to the Mt. Pinatubo Eruption

[5] The eruption of Mt Pinatubo in 1991 provided a powerful natural experiment as a transient perturbation to the climate system. Advances in remote sensing mean that this event was better observed, particularly from space, than previous eruptions. Several studies have used the eruption to examine the radiative response and recovery of the planet following this perturbation. For example, some workers have attempted to use the observed response of near surface temperature following the Pinatubo eruption (and other volcanic eruptions) to determine the climate sensitivity [e.g., *Lindzen and Giannitsis*, 1998; *Douglass and Knox*, 2005; *Wigley et al.*, 2005], with varying results.

[6] *Soden et al.* [2002] used the eruption as a perturbation in an elegant study of the water feedback in climate. These authors showed that model simulations of the humidity, temperature and radiative responses of the atmosphere to the perturbation could only be adequately modeled if the water vapour feedback was included, providing evidence in support of a positive water vapour feedback. Building on this work, *Forster and Collins* [2004] used the response to the eruption as simulated in the Hadley Centre model, and as observed in surface and satellite measurements to attempt to directly estimate the magnitude of the water vapour feedback parameter.

[7] An examination of the data presented by *Soden et al.* [2002] suggested another use of this natural experiment, this time to study the natural response times of feedback processes in the climate system. It was apparent that information existed in their analysis on the time constants associated with various processes, both dynamical and radiative, associated with the Earth's response to the Pinatubo perturbation, including new information on the variability of F_N at the TOA.

[8] Here we use the time series of a variety of measured parameters published by *Soden et al.* [2002]. Their original radiation budget data (Edition 2 ERBS fluxes) were kindly made available to us by the authors. However, in response to a reviewer's helpful suggestion we have instead used the corrected Edition 3, revision 1 version of this dataset, using 72 day means to avoid aliasing of diurnal variability onto long term trends [*Wong et al.*, 2006]. Figure 1 presents time series of the anomalies of the following parameters:

[9] 1. Observed longwave and shortwave TOA fluxes for latitudes 60N–60S and for 1991–1996 (note: the latitude range was chosen because this was the maximum extent of the observations used from ERBS (see http://asd-www.larc.nasa.gov/erbe/erbs.html): however, *Soden et al.* [2002] confirmed that there were no significant differences in the model when averaged from 60N–60S and from 90N–90S).

[10] 2. The net flux anomaly, formed from the difference between absorbed SW and emitted LW fluxes.

[11] 3. Observed total column water vapour, and lower tropospheric temperature, for 90N–90S and same time period taken from the NVAP project (NASA Water Vapor Project, see http://eosweb.larc.nasa.gov/PRODOCS/nvap/table_nvap.html [*Randel et al.*, 1996]).

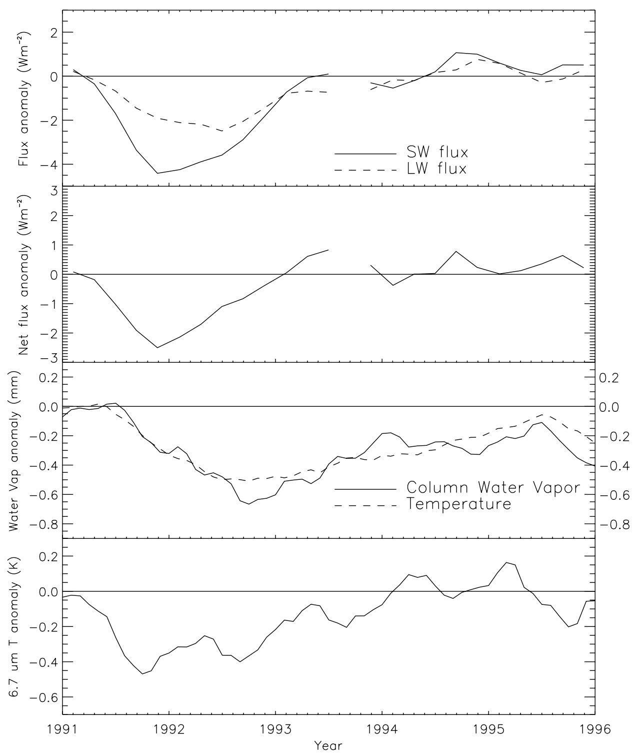

Figure 1. The time series of the anomalies of the following parameters [adapted from *Soden et al.*, 2002]: (top to bottom) observed longwave and shortwave TOA fluxes for latitudes 60N–60S and for 1991–1996, observed net flux formed from the difference between absorbed SW and emitted LW fluxes, observed total column water vapour and lower tropospheric temperature for 90N–90S and same time period, and observed 6.7 μm brightness temperature for 90N–90S and same time period.

[12] 4. Observed 6.7 μm brightness temperature for 90N–90S and same time period TOVS Radiances Pathfinder project (NOAA TIROS Operational Vertical Sounder, see http://www.ozonelayer.noaa.gov/action/tovs.htm [*Bates et al.*, 1996]).

[13] Using these time series, we have determined the growth and decay times constants of the response to the eruption seen in each of these fields, ie in the TOA fluxes (LW, SW and net flux, F_N), in tropospheric humidity and temperature, and in the 6.7 μm brightness temperature. We make the assumption that these time dependences are exponential, and fit the observations by taking logarithms and fitting straight lines separately to the growth and the decay parts of the curves in Figure 1. The results for the e^{-1} time constant associated with the growth and decay responses, along with values of the peak in each anomaly parameter, and the timing of that peak, are given below in Table 1.

3. Analysis of the Derived Time Constants

3.1. Broadband Fluxes

[14] From Figure 1 we see that the absorbed SW flux anomaly responds almost immediately after the eruption and falls quickly to a minimum at -4.4 W m^{-2} in October

Table 1. Exponential Growth and Decay Time Constants Determined From Time Series in Figure 1[a]

Parameter	Growth Time Constant, τ_G, months	Decay Time Constant, τ_D, months	Peak Anomaly Time	Peak Anomaly Magnitude
Absorbed SW flux anomaly, W m^{-2} (60N–60S: ERBE data)	2–4	9–11	Oct 1991	−4.2 W m^{-2}
Emitted LW flux anomaly, W m^{-2} (60N–60S: ERBE data)	3–5 (but lasts longer than SW before decaying)	6–8	April 1992	−2.1 W m^{-2}
Tropospheric total column water vapour anomaly, mm (90N–90S: NVAP data)	6–10	23–31	July–Oct 1992	−0.6 to −0.8 mm
Lower tropospheric temperature anomaly, K (90N–90S: MSU data)	6–10	23–31	April–Sept 1992	−0.5 to −0.7 K
6.7 μm brightness temperature (90N–90S: HIRS data)	2–3	9–13	Nov 1991 (but very flat till Sept 1992)	−0.4 to −0.5 K

[a]Pinatubo erupted in June 1991. Time constants for the net flux anomaly (F_N) are the same, within measurement and fitting errors, as those for the SW anomaly.

1991, decays back towards zero anomaly, overshooting slightly, before relaxing back towards zero. The peak negative anomaly occurs quickly, only 4 months, after the eruption, presumably due to the rapid introduction of aerosols into the stratosphere, which directly scatter sunlight back to space [*Self et al.*, 1999].

[15] The LW anomaly peaks at about −2.5 W m^{-2}, that is a reduction in emission, with a rise time slightly longer than the SW, presumably because the atmospheric aerosol clears from the atmosphere faster than the thermal inertia of the Earth allows the growth of the emission anomaly. After the peak, the LW anomaly returns to zero by early 1994, and shows a small ($\leq$ 1 W m^{-2}) positive anomaly by early 1995, slightly lagging the SW 'overshoot'.

[16] The peak difference between the SW and LW fluxes gives a peak change in net radiation anomaly, F_N, of about −2.6 W m^{-2} in October 1991, a loss of energy to the planet that cools the surface and the lower atmosphere. We estimate a relative uncertainty in this net flux anomaly of about ±0.5 W m^{-2}. This value of F_N compares well with the value of −2.7 ± 1.0 W m^{-2} found for the global mean net flux anomaly found by *Minnis et al.* [1993] also using ERBE non-scanner data. *Wielicki et al.* [2002] found a peak change in net radiation anomaly of about −7 W m^{-2} for the 20N–20S zone (subsequently corrected to closer to −6 W m^{-2} (B. A. Wielicki private communication, 2006)). This difference between low latitudes and the global average arises because the volcano erupted in the tropics, and as observations from a number of satellites indicated [e.g., *Holasek et al.*, 1996], the aerosol was initially highly concentrated at low latitudes, subsequently extending to higher latitudes with time and thinning. The observed fluxes will, of course, include all significant effects, including the radiative impacts of the observed decrease in stratospheric ozone [*Schoeberl et al.*, 1993] and the increase in methane [*Dlugokencky et al.*, 1996] observed following the eruption, both of which will make a small contribution to the observed fluxes, acting to reinforce the direct effects of the aerosols on the SW and LW fluxes respectively.

3.2. Humidity and Temperature

[17] The troposphere responded to this loss of net radiation by cooling (by about 0.5 K), especially the lower troposphere, and by a decrease of total column water vapour (by about 0.6 mm). Note that the onset of the temperature and humidity responses are a little delayed following the radiative perturbation. The growth of these anomalies is slower than for the radiative fluxes (6–10 months rather than 2–5 months), presumably because the growth of the humidity anomaly depends on dynamical processes and heating of the bulk troposphere, and not just on the very rapid arrival of aerosols in the atmosphere. Indeed, a timescale of several months must involve the ocean. The result is similar for the decay phase, with time constants of 23–31 months compared with the 6–11 months for the SW and LW fluxes. These results are consistent with the findings of *Soden et al.* [2002], who showed that model simulations of the Pinatubo event could only produce the correct bulk property variations with time if the water vapour feedback was enabled, involving as it does longer time constants associated with bulk evaporation and transport processes.

3.3. 6.7 μm Brightness Temperature

[18] The 6.7 μm brightness temperature is, by design, sensitive to water vapour in the 200–500 hPa layer, and relatively insensitive to temperature: this ensures maximum sensitivity to water vapour in this layer with minimum sensitivity to temperature errors. This parameter shows an initial rapid reduction, significantly faster (2–3 months) than the atmospheric humidity, and similar to the radiative fluxes, with a decay time constant of 9–13 months, which is also closer to those of the fluxes than the atmospheric parameters. This is discussed further below. There is also a conspicuous annual pattern in the 6.7 μm brightness temperature.

4. Discussion

[19] The overall behaviour or the various parameters is consistent with a rapid SW response to the volcanic injecta from Mt. Pinatubo, and a slightly delayed response in the LW anomaly that accompanies it, presumably due to the thermal inertia of the system. The responses of the bulk fields of temperature and humidity are considerably slower, because of the longer time scales of the dynamical and thermodynamic processes that are involved in the response of these fields to the radiative perturbation (e.g., heating, evaporation, dynamics). The 6.7 μm brightness temperature anomaly, which depends on both radiative and dynamical/

thermodynamic processes shows time constants closer to the radiative parameters, for reasons explored below.

[20] The time constants we have derived (Table 1) for these various components of this event are consistent with previous process studies. For example, the time constant for the onset of the radiative components of the event, about 2–5 months, is consistent with the results of *Minnis et al.* [1993], who found a maximum aerosol loading in September 1991, just 3 months after the eruption, while *Stenchikov et al.* [1998] suggest the maximum visible optical depth is found in October to November, some 4–5 months after the eruption. The decay time constant back to recovery of radiative 'equilibrium' of $\sim$ 6–11 months, is close to the time constant for the removal of volcanic aerosol from the stratosphere, which *Kent and Hansen* [1998] estimated at $\sim$ 9.5 months using lidar observations.

[21] Certain questions remain, or arise as a result of this work. The first is the question of the net flux, F_N. Here the question concerns evidence for departures from $F_N = 0$; how well do the TOA fluxes balance, and what timescales are involved in variations? In other words, do we anticipate that the Earth remains in, or close to, radiative balance, and if so, with what time constant? The results shown here for F_N are dominated by the SW term. We see a rapid fall to about -2.5 W m^{-2}, then a return to very close to zero by the beginning of 1993. The value then remains close to zero (or a very slight positive anomaly) for the remainder of the timeseries. This accords with *Wielicki et al.* [2002], who show that in the tropics the net flux departs from zero only for the duration of the SW anomaly, and tracks this anomaly quickly back to zero (and thereafter remains remarkably close to zero, at least within the 'noise' of the results). The model studies of *Hansen et al.* [2005] seem to indicate a background global net flux imbalance of $F_N \simeq +1$ W m^{-2} by the year 2000, rising from zero in about 1960, caused by the thermal inertia of the oceans slowing the temperature response to the extra heat in the climate system caused by increasing greenhouse gases. Thus, the data we do have would imply a rapid recovery in TOA net flux to zero following any perturbation, without any significant lag associated with thermal inertia, but the noisy nature of the data does not allow the latter to be ruled out. Further work on the net balance of radiative energy at the TOA is needed.

[22] A second question is, can we describe the Earth as being in a definable steady state, and if so, with what variability? The thermodynamic system represented by the Earth, even if in energy balance at the TOA, is driven by the production of entropy that arises due to many scales of processes between the incoming solar radiation and the outgoing infrared, and is not in equilibrium. A future paper will discuss this more general question of the thermodynamics of the Earth and its climate system.

[23] Another question of interest that arises from these results concerns the 6.7 μm brightness temperature anomaly. Since the instrumental parameters for this channel have been purposely selected to be sensitive to water vapour amount in the 200–500 hPa layer, and relatively insensitive to temperature (both requirements for maximum sensitivity to humidity and minimum sensitivity to anything else), it is perhaps surprising that the response times for T6.7 (Table 1) are not closer to those for the humidity response. The answer to this may involve a compensation effect on outgoing LW radiation between the temperature lapse rate and humidity in the free troposphere [see, e.g., *Held and Soden*, 2000]. Thus, if an increase of humidity occurs at some level in the atmosphere, then emission to space from that layer increases. However, the increased opacity associated with the increase in water vapour concentration causes the mean level of emission to rise to a colder layer (because of the temperature lapse rate) and so the emission to space is reduced. Thus the direct effect of increasing water vapour amount is compensated by a shift to a lower emission temperature, leaving the direct radiative effect of the aerosols to dominate the 6.7 μm channel recovery.

[24] In conclusion, studying the responses of the radiative, humidity and temperature fields has allowed us to address the separation of different processes which act with different characteristic time constants. It will be valuable as an extra test of fidelity to ensure that climate models predict the correct rates of change for such processes. This study has also allowed us to examine the effect of a perturbation on the top of atmosphere net radiative flux. The rapid restoration of zero net flux at the TOA found in this and other studies raises a question over the utility of ERB measurements on a global scale for climate studies: ERB measurements are, however, of critical value in studies of more localised (eg regional) processes in the climate system.

[25] **Acknowledgments.** We are grateful to Brian Soden for providing the data to us contained in the figures from his paper [*Soden et al.*, 2002], and to two anonymous reviewers for their helpful comments on the data and thermodynamics of the Earth system. Edition 3, revision 1 ERBS data was obtained from the NASA Langley Data Center, at http://earth-www.larc.nasa.gov/erbeweb/Edition3_Rev1/.

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