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**Stratospheric
humidity and climate
sensitivity**

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Stratospheric water vapour and high climate sensitivity in a version of the HadSM3 climate model

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Abstract

It has been shown previously that one member of the Met Office Hadley Centre single-parameter perturbed physics ensemble – the so-called “low entrainment parameter” member – has a much higher climate sensitivity than other individual parameter perturbations. Here we show that the concentration of stratospheric water vapour in this member is over three times higher than observations, and, more importantly for climate sensitivity, increases significantly when climate warms. The large surface temperature response of this ensemble member is more consistent with a feedback associated with the stratospheric humidity change, rather than high clouds as has been previously suggested. The direct relationship between the bias in the control state (elevated stratospheric humidity) and the cause of the high climate sensitivity (a further increase in stratospheric humidity) lends further doubt as to the realism of this particular integration. This, together with other evidence, lowers the likelihood that the climate system’s physical sensitivity might be significantly higher than the likely upper range quoted in the Intergovernmental Panel on Climate Change’s fourth assessment report.

1 Introduction

Much discussion has centred on the likelihood of the sensitivity of the climate system being significantly larger than the 2–4.5 K range quoted in the Intergovernmental Panel on Climate Change (IPCC)’s fourth assessment report (AR4) (IPCC, 2007). There are various lines of evidence that support the possibility of high climate sensitivities; one has focused on the results of a set of experiments carried out using the Met Office Hadley Centre’s HadSM3 climate model.

The Quantifying Uncertainty in Model Prediction (QUMP) ensemble (Murphy et al., 2004) consisted of a series of general circulation model or GCM integrations with different perturbed parameters representing in some senses uncertainties in these processes. The integration that is the subject of this paper is the so-called low entrainment

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parameter (hence LEP) integration. When entrainment rates in the model's convection scheme are set to low values, the climate sensitivity is approximately 7 K on doubling CO₂ from pre-industrial values, which is much higher than the IPCC range of 2–4.5 K quoted above, and much higher than any other member of the single-parameter Murphy et al. (2004) ensemble.

It is clearly important to assess the validity of the LEP run, given that such a high sensitivity would have profound implications for climate change in the latter half of the 21st century and beyond, given current emissions projections, and an equivalently profound impact on international negotiations to limit emissions. Such analyses have been carried out, and while this member's climate is in some senses further away from observations than other members of the QUMP ensemble, it cannot be ruled out (Murphy et al., 2004; Stainforth et al., 2005). Here we focus on one aspect of the LEP run: its high stratospheric humidity, and the implications of changes in this quantity for the validity of the LEP run, and the feedback processes occurring in it.

Elevated values of humidity in the upper tropospheric/lower stratospheric (UTLS) region have been noticed before by Sanderson et al. (2008). They found relative humidity (RH) changed by 30% on doubling CO₂ in a version of the LEP run carried out by the Climateprediction.net project (Stainforth et al., 2005). They inferred that high cloud in the UTLS region was responsible for the high sensitivity. However, their Fig. 5 shows high values of RH in the tropics at the 20–25 km level compared to a control simulation, which is not only at a much higher altitude than the cold point of the tropical tropopause, but also insufficient to cause cloud formation in such a dry region. This study explores an alternative interpretation – that stratospheric water vapour (henceforth SWV) changes rather than cloud changes are the main cause of the high climate sensitivity of the LEP run.

In a standard HadSM3 simulation, water vapour is freeze dried as it reaches the coldest point of the tropical tropopause; this leads to very low values of SWV of approximately 2–3 ppmv, consistent with observations. High values of SWV are seen in the LEP run because less entrainment in convection reduces the dilution of convective

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plumes by dry air. The plumes are therefore more intense, and cause the upper tropical troposphere to moisten far more than in the standard simulation. The moister air is then available for transport from the upper troposphere into the lower stratosphere isentropically in the subtropics, where the tropopause height changes rapidly, and isentropes cross the tropopause. Such transport has actually been noticed previously in the HadCM2 model (D. Karoly, personal communication, 2009).

In this paper we show that SWV biases in the LEP run are far worse than suggested by Sanderson et al. (2008), and cast doubt on the plausibility of this ensemble member's climatology. We then show that the extra radiative effect associated with the stratospheric moisture change in the $2 \cdot \text{CO}_2$ LEP integration is almost as large as the CO_2 forcing itself, and can explain the high climate sensitivity of LEP, and rule out cloud changes as a substantial contributor to the differences in sensitivity between the LEP and the standard version of HadSM3. We then discuss our results in the context of constraining climate sensitivity.

2 Results

We present results from four integrations of the HadSM3 model: a standard-parameter control run and an LEP run with pre-industrial CO_2 (STD1 and LEP1, respectively); a standard-parameter control run and an LEP run with $2 \cdot$ pre-industrial CO_2 (STD2 and LEP2, respectively).

STD1 and STD2 exhibit values of q in the stratosphere broadly consistent with observations (not shown). The change in q between STD1 and STD2 under enhanced CO_2 is small. Figure 1 shows the stratospheric specific humidity, q , in LEP1 and LEP2. LEP1 exhibits values of q that are much higher than the observed mean values of approximately 4 ppmv (e.g. Rosenlof et al., 2001), and which occur throughout the stratosphere. The large hemispheric asymmetry also appears inconsistent with observations. Sanderson et al. (2008) suggested that the differences between LEP1 and STD1 are concentrated in the UTLS region. We suggest that the reason for their

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interpretation is that they diagnosed differences in RH rather than q : the choice of the former magnifies differences where RH is large, i.e. near the cold point of the tropical tropopause at the 100 hPa level. Consider two levels having similar values of q , but RH values of 1% and 25. If q is then doubled at both levels, the former level will exhibit a change in RH of 1%, whereas the latter will show a change of 25%, which underestimates the importance of mid-stratospheric changes.

LEP2 (Fig. 1 grey dashed line) has values approaching 40 ppmv in the mid-stratosphere, which is an order of magnitude higher than present-day observations. LEP2 exhibits positive anomalies in the subtropics, which is where the tropopause drops in height, and isentropes can cross it. These anomalies are consistent with humid air in LEP2 being isentropically transported polewards from the troposphere into the lower stratosphere, and being uplifted in the Brewer-Dobson circulation.

One can confirm the radiative importance of the water vapour in LEP1 by analysing the energy budget in terms of downward short wave (SW) and long wave (LW) radiation at the tropopause in runs STD1 and LEP1. The LW difference is $+1.2 \text{ Wm}^{-2}$, whereas the SW difference is only -0.1 Wm^{-2} , showing that LW effects arising from the difference in water vapour are dominating the difference in downward radiation at the tropopause between STD1 and LEP1.

The difference in downward LW flux at the tropopause between STD2 and STD1 is 0.6 Wm^{-2} , which can be largely attributed to the radiative effects of more CO_2 in the stratosphere. There is no difference in downward SW flux. However, the difference in downward tropopause LW flux between LEP2 and LEP1 is 3.3 Wm^{-2} , while the difference in downward SW flux is 0.1 Wm^{-2} , suggesting that the extra stratospheric humidity (and cooling associated with the extra humidity) in LEP2 is contributing 2.8 Wm^{-2} to the radiative budget after doubling CO_2 compared to run STD2.

We have attempted to confirm that the extra radiative effect is associated with the extra SWV in LEP2 by three means. Firstly, Fig. 2 shows the timescale over which both the SWV anomaly and downward LW forcing at the tropopause build up. The solid curves in Fig. 2 (top) corresponding to STD1 and STD2 show negligible trends.

However, run LEP2, shown by the dashed grey line, exhibits an increase in stratospheric humidity over the first 10 years of the integration. The dashed grey curve in Fig. 2 (bottom) shows how the downward LW flux at the tropopause evolves in response to the humidity anomaly in LEP2: it too increases over a timescale of 10 years until equilibrating at a value of 3.3 Wm^{-2} above the LEP1 value, suggesting it is associated with the SWV anomaly.

As a second test of our hypothesis, we have calculated the radiative forcing at the tropopause resulting from a uniform change in SWV from 10 ppmv to 20 ppmv (the approximate mean SWV concentrations of the LEP1 and LEP2 integrations), using the fixed-dynamical-heating or FDH approach (e.g. Forster and Shine, 2002), in the HadSM3 radiation code. The FDH forcing is 2.7 Wm^{-2} , which is very close to the 2.8 Wm^{-2} additional downward LW flux at the tropopause between LEP2 and LEP1 compared to STD1 and STD2. This shows that the extra SWV in LEP2 is capable of explaining a large component of the extra downward LW forcing in that run.

Finally, we have estimated what the climate sensitivity would be for the STD and LEP experiments if their clear-sky and cloud feedback parameters were interchanged. Following the method of Webb et al. (2006), and assuming a standard HadCM3 value for CO_2 forcing of 3.75 Wm^{-2} for both experiments, the clear-sky feedback parameters for STD and LEP are -1.33 and $-0.79 \text{ Wm}^{-2} \text{ K}^{-1}$, respectively, while the cloud feedback parameters are 0.21 and $0.24 \text{ Wm}^{-2} \text{ K}^{-1}$. Estimating the climate sensitivities from the feedback parameters yields 3.3 and 6.8 K for STD and LEP respectively. The STD clear-sky feedback combined with the LEP cloud feedback yields 3.4 K , while the LEP clear-sky feedback combined with the STD cloud feedback yields 6.5 K . Hence the difference in the clear-sky feedback between the STD and LEP experiments explains 95% of the difference in their climate sensitivities.

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The radiative forcing associated with doubling CO_2 from pre-industrial concentrations (in HadCM3) is 3.7 Wm^{-2} . If the extra downward LW effect associated with SWV in the LEP2 experiment is 2.7 Wm^{-2} , this will almost double the total radiative forcing.

5 The effects of the extra SWV therefore explain the high sensitivity of the LEP1/2 model incarnation. Our results suggest that the tropospheric feedbacks in LEP1/2 are similar to other members of the Murphy et al. (2004) ensemble, all of which have a much lower temperature response.

One can answer the question of whether the stratospheric water vapour response in LEP2 is an indirect forcing or a feedback (the latter being dependent on surface change) by plotting the evolution of the temperature at 1.5 m vs. the top-of-atmosphere (hence TOA) net flux in run LEP2, in the manner of Gregory et al. (2004). In their analysis, points lie along more or less a straight line with a negative gradient as the temperature warms and TOA flux reduces to zero. Figure 3 shows that in the first 10 5–10 years of model integration, when Fig. 2 shows that SWV is increasing in LEP2, TOA flux actually increases, before decreasing in line with Gregory et al. (2004). This implies that the SWV response is a feedback, and is taking place on a much longer timescale than might be implied by a change in cirrus cloud in the UTLS.

Different methods have been used to assess the likelihood of the climate system's sensitivity mirroring the magnitude of the LEP1/2 system; some have been based on comparing the climatology of individual ensemble members with time-averaged observations (Murphy et al., 2004; Collins et al., 2009) while others use novel tests using different numerical weather prediction models (Rodwell and Palmer, 2006). The key difference in the present work is that the process causing the large stratospheric humidity bias in LEP1 appears to be the same process that is responsible for the water vapour increase, and hence the large temperature response, in LEP2. There is therefore a stronger case for considering the temperature response in LEP2 to be implausible.

25 A scenario that should be considered is whether the high temperature response in

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LEP2 might occur in reality because of a real change in convective entrainment or other processes that significantly increase SWV in a warmer climate. There has indeed been an increasing trend in stratospheric humidity over the latter half of the 20th century, but this trend is very noisy (e.g. Rosenlof et al., 2001), has many possible causes not related to climate warming (e.g. Scaife et al., 2003; Joshi and Shine, 2003), and at present is hard to attribute (Fueglistaler and Haynes, 2005). The trend has actually been zero since the year 2000 (Randel et al., 2006). Additionally, since LEP2 exhibits a radiative effect from the change in SWV that is about 80% of the CO₂ forcing, one might expect that the radiative forcing associated with observed SWV changes since pre-industrial times should be a significant fraction of the 1.6 Wm⁻² associated with CO₂ since 1860, if the real world behaved like LEP. Forster and Shine (2002) estimated a value of only 0.29 Wm⁻² for stratospheric water forcing in the 20th century, but this was based on the peak trend, which has now lessened.

Figure 4 shows the sensitivity of q at 50 hPa to 1.5 m temperature in the LEP2 run. The gradient is approximately 3.7 ppmv K⁻¹ during the transient phase; if such a feedback had happened in the 20th century, when globally averaged temperatures rose by 0.8 K, q should have increased by almost 3 ppmv, which is much higher than the observed trend (see above and Rosenlof et al., 2001). We conclude that it is therefore highly unlikely that the observed trend in SWV is consistent with the LEP1/LEP2 integrations, although some SWV feedback of this nature, albeit having a much smaller magnitude, might operate under enhanced levels of CO₂. Further work is required on this topic.

Future research in this area should involve examining the response of the HadSM3 model when multiple parameters are perturbed at the same time, given the known interaction of the low entrainment parameter with other perturbations (Rougier et al., 2010). The robustness of our results to multiple parameter perturbations could also be quantified in this way. For example, Rougier et al. (2010) show that relatively large values of climate sensitivity are possible in HadSM3 for much more reasonable values of the entrainment parameter.

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4 Conclusions

We have investigated the “low-entrainment-value” parameter pre-industrial and 2 · CO₂ climates of the HadSM3 ensemble. We find that the high sensitivity of this climate is because of a large increase in stratospheric water vapour in the 2 · CO₂ integration. Given that this is a result of a process that also causes a very large bias in the stratospheric humidity in the present-day climate, it is very unlikely that the real climate system has a sensitivity this high for this reason.

This analysis has again shown that changes to minor constituents in the stratosphere can have profound effects on the evolution of the surface climate in models. Any future metrics of model behaviour should take account of potential biases arising from this region of the atmosphere.

Finally, we do note that although it is likely that the climate system as represented by HadSM3 does not have a high sensitivity, our results say nothing about the sensitivity of the full Earth system, when feedbacks not in HadSM3 such as the carbon cycle are taken into account (e.g. Friedlingstein et al., 2006). It is entirely possible that such feedbacks add significantly to the temperature response of the Earth system for a given radiative forcing. Further research should be done on constraining these sorts of Earth-system-type sensitivities.

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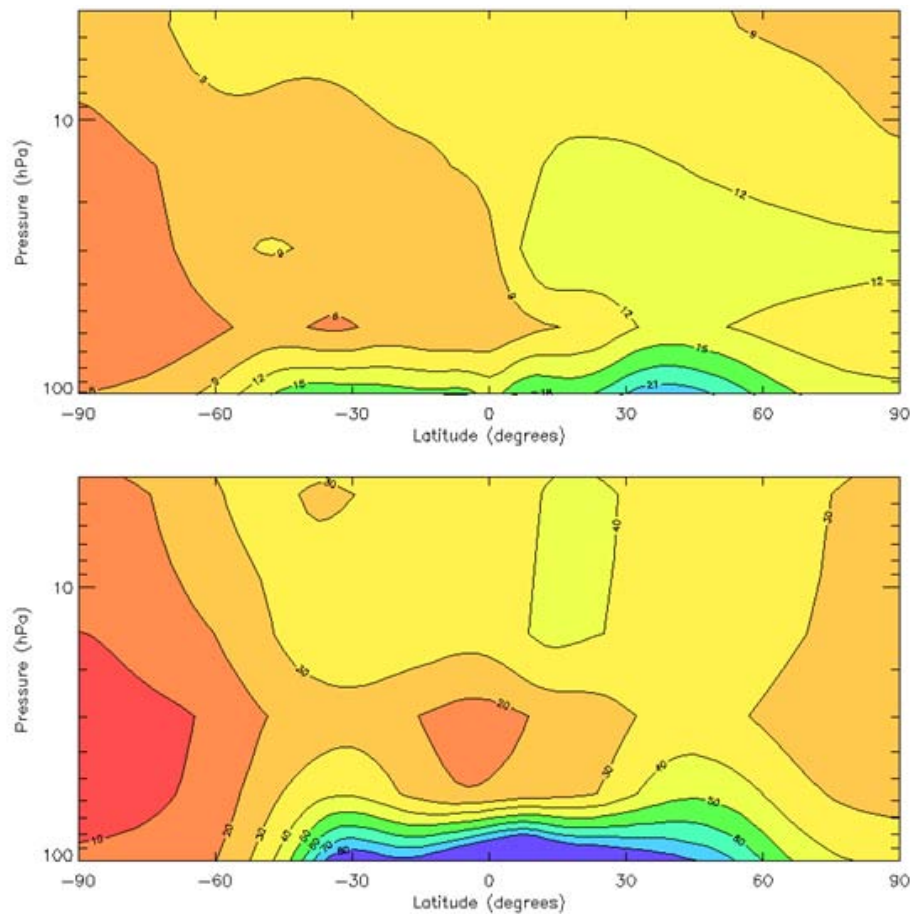


Fig. 1. Time averaged zonal cross sections of specific humidity q in LEP1 (top) and LEP2 (bottom). Note the different contour intervals in the two plots (3 ppmv and 10 ppmv, respectively).

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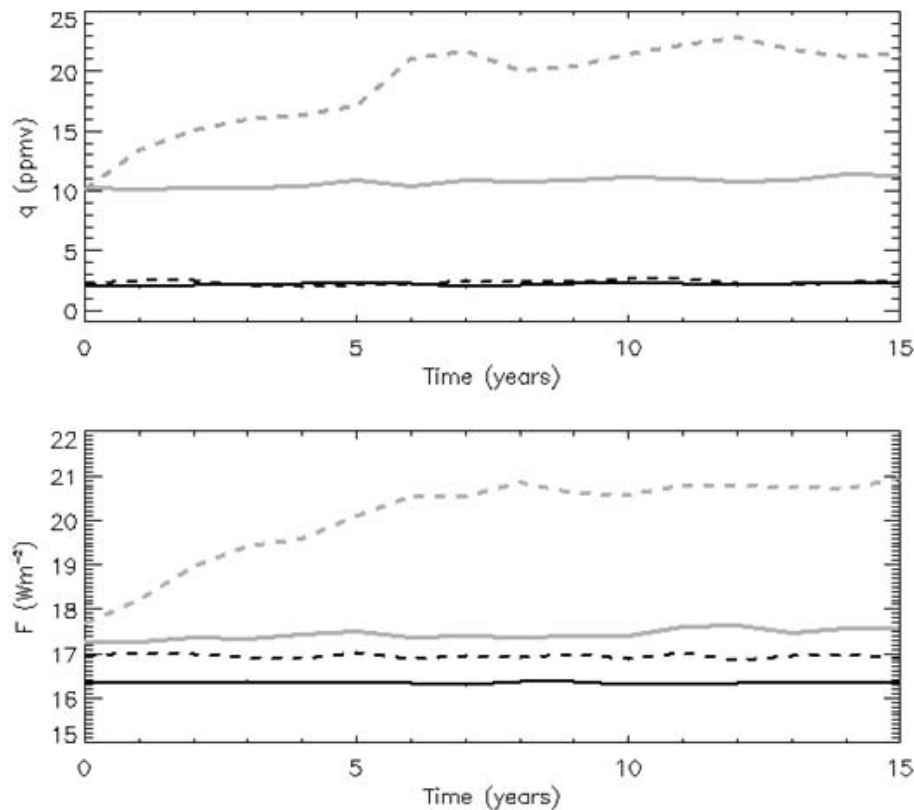


Fig. 2. Top panel: the evolution of globally averaged specific humidity in time in STD1 (solid black); STD2 (dashed black); LEP1 (solid grey) and LEP2 (dashed grey). Bottom panel: as for the evolution of downward LW radiation at the tropopause.

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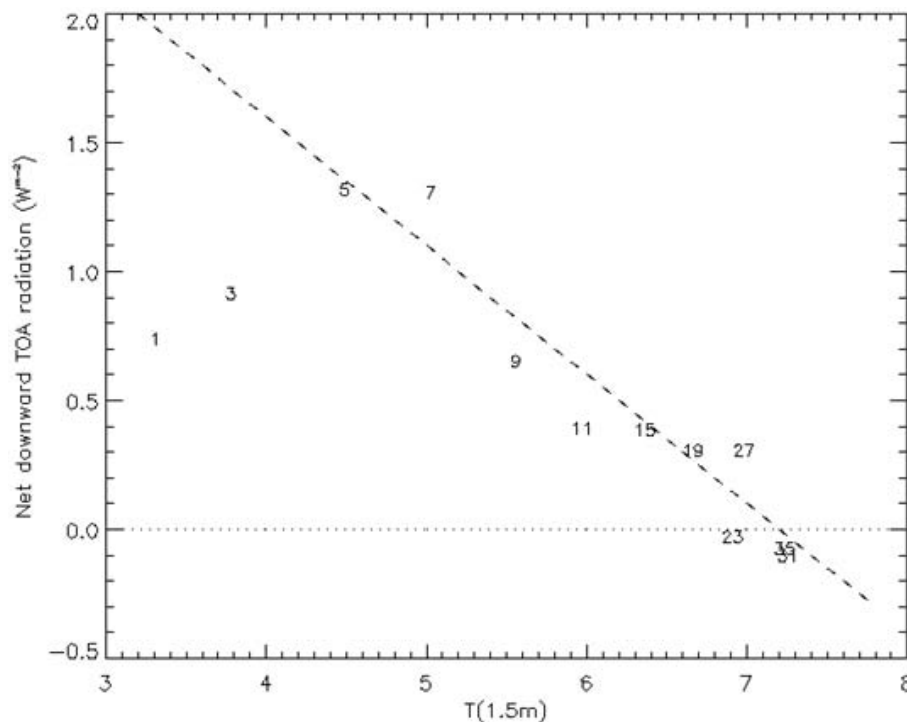


Fig. 3. Anomalous net top-of-atmosphere downward flux in LEP2 vs. surface temperature change during the transient phase of the integration. Each axis has had the mean value for that quantity in run LEP1 subtracted from it. Each number corresponds to the average year of the integration. Years 1–10 have biannual means plotted, while years 10–35 have quadrennial means plotted. The dashed line corresponds to the linear regression $\text{TOA} = 3.6 - 0.5 T$.

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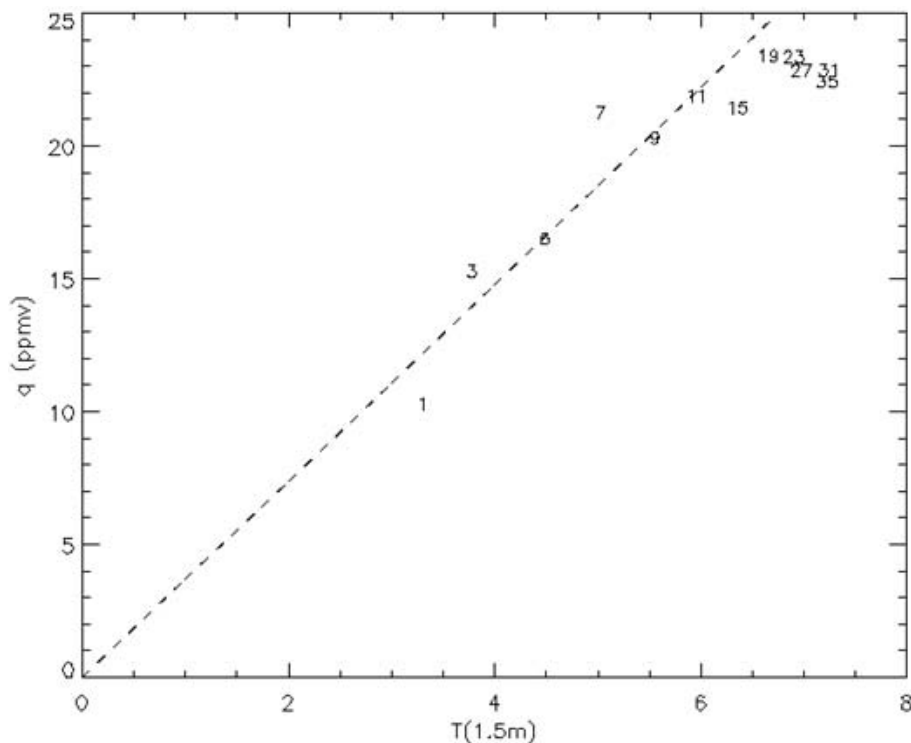


Fig. 4. q at 50 hPa in LEP2 vs. surface temperature during the transient phase of the integration. The x-axis has had the mean temperature in run LEP1 subtracted from it. The numbers are calculated as in Fig. 3. The dashed line corresponds to the linear regression $q=3.7T$.

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