Joint Dependence of Longwave Feedback on Surface Temperature and Relative Humidity

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Abstract Various studies have suggested that Earth’s clear-sky outgoing longwave radiation (OLR) varies linearly with surface temperature, with a longwave clear-sky feedback that is, independent of surface temperature and relative humidity. However, this uniformity conflicts with the notion that humidity controls tropical stability (e.g., the “furnace” and “radiator fins” of Pierrehumbert (1995, https://doi.org/10.1175/1520-0469(1995)052%3C1784:TRFATL%3E2.0.CO;2)). Here, we use a column model to explore the dependence of longwave clear-sky feedback on both surface temperature and relative humidity. We find that a strong humidity dependence in the feedback emerges above 275 K, which stems from the closing of the H2O window, and that the furnace and radiator fins are consequences of this dependence. We then clarify that radiator fins are better characterized by tropical variations in clear-sky feedback than OLR. Finally, we construct a simple model for estimating the all-sky feedback and find that although clouds lower the magnitude of longwave feedback, the humidity-dependence persists.

Plain Language Summary The dependence of outgoing longwave radiation radiation on surface temperature (i.e., the feedback) is a major determinant of the climate's stability. Various studies have suggested that the feedback is largely independent of both surface temperature and relative humidity, which implies that the climate stability is also independent of surface temperature and relative humidity. However, this uniformity seems to contradict other work which shows that the subtropics are relatively stable and the deep tropics are relatively unstable, implying the feedback must vary between the two regions. We resolve this apparent contradiction by systematically computing the feedback as a function of both surface temperature and relative humidity. Above 275 K, the feedback depends significantly on relative humidity. We then show the feedback does indeed vary in the tropics and that this difference arises from regional differences in relative humidity. Finally, we estimate the effects of clouds on the feedback with a simple model and find that although clouds have a destabilizing influence, the significant dependence on relative humidity persists. Our work gives a renewed appreciation for how the feedback can vary significantly with both surface temperature and relative humidity.

1. Introduction

The longwave clear-sky feedback parameter $\lambda_s$ relates a change in clear-sky outgoing longwave radiation $OLR_s$ to a change in surface temperature $T_s$,

$$\lambda_s = \frac{dOLR_s}{dT_s} \left[\text{Wm}^{-2}\text{K}^{-1}\right].$$

(1)

It is a measure of the stability of the climate and thus is a well studied quantity, with a canonical value for its global mean of about 2.2 ± 10% Wm⁻²K⁻¹ (Allan et al., 1999; Bony et al., 1995; Budyko, 1969; Cess et al., 1989, 1990; Chung et al., 2010; Dessler et al., 2008; Jeevanjee, 2018; Koll & Cronin, 2018; Raval et al., 1994; Zhang et al., 2020).

The convergence of the global mean value of $\lambda_s$ across both observations and the model hierarchy suggests robust physics that is, insensitive to the idiosyncrasies of the individual studies. Recently, Koll and Cronin (2018) gave an explanation of this physics as a balance between increasing surface Planck feedback and decreasing surface transmissivity. They verified that $\lambda_s \approx 2.2$ Wm⁻²K⁻¹ for a wide range of $T_s$ in a column model. Zhang et al. (2020) then extended this analysis to GCMs and similarly found $\lambda_s$ to be independent of both $T_s$ and free-tropospheric relative humidity (RH).
This work on the uniformity of feedback lies in tension with the notion that meridional variations in clear-sky relative humidity are important in controlling tropical stability. In particular, Pierrehumbert (1995) argued that the warm and moist deep tropics, with active deep convection (furnace) are close to a local runaway greenhouse, but are radiatively stabilized by the warm, yet dryer, and more quiescent subtropics (radiator fins). However, Pierrehumbert (1995) was equivocal on whether the furnace and radiator fins manifest as tropical variations in OLR\textsubscript{cs}, or rather in \( \lambda_{cs} \), which is the more relevant parameter for stability. Indeed, as we shall show later, the latitudinal variations in OLR\textsubscript{cs} within the tropics are quite muted compared to OLR\text sub {cs} variations over the globe. Here, then, we will pursue the idea that radiator fins manifest instead as tropical variations in \( \lambda_{cs} \).

Clouds are another process that may play a role in controlling the structure of zonal-mean feedback. Pierrehumbert (1995) argued for the presence of tropical furnaces and radiator fins using only clear-sky physics. However, humid regions and cloudy regions often go hand-in-hand, and high clouds are known to have a robust influence on the longwave feedback (Zelinka & Hartmann, 2010), so we might expect the longwave all-sky feedback parameter \( \lambda_{cs} \) to look different from \( \lambda_{as} \) in the zonal mean.

We lack clarity on whether the taxonomy of the furnace and radiator fins are better described by OLR\textsub {cs} or by \( \lambda_{as} \). There is also a tension between the constancy of \( \lambda_{as} \) observed across studies and the notion that humidity variations control tropical stability, and it is unclear how clouds might modulate this relationship. This state of affairs motivates us to ask the following questions:

1. Do furnaces and radiator fins indeed manifest as a contrast in the zonal-mean \( \lambda_{as} \) as opposed to the zonal-mean OLR\textsub {cs}?
2. How do we reconcile variations in \( \lambda_{as} \) implied by furnaces and radiator fins when other studies suggest \( \lambda_{as} \) is approximately constant?
3. How do clouds modify the meridional structure of longwave feedback?

To this end we first construct a “phase space,” in which \( \lambda_{as} \) is computed as a joint function of \( T_s \) and column RH. Below 275 K, we find that \( \lambda_{as} \) stays within 10% of 2.2 W m\textsuperscript{-2} K\textsuperscript{-1}, even as RH varies. Above 275 K, however, a significant RH-dependence emerges, leading to much greater variations in \( \lambda_{cs} \). We show that this RH-dependence stems from the closing of the H\textsubscript{2}O window. The tropical contrast in zonal-mean \( \lambda_{as} \) is, then, a consequence of this RH-dependence at high temperatures. Finally, we construct a simple model for evaluating the all-sky feedback and find that although clouds decrease the zonal-mean feedback, the RH-dependence remains significant.

### 2. Results

#### 2.1. Exploring the State Dependence of \( \lambda_{as} \)

We first address Question 2 by exploring the state dependence of \( \lambda_{as} \) as a function of both \( T_s \) and RH. We use RH as a state variable because RH-based feedbacks have certain advantages over specific humidity (\( q_v \)) based feedbacks both from a thermodynamic point of view (Held & Shell, 2012) and from a radiative point of view (Jeevanjee et al., 2021), as specific humidity already has a de facto strong temperature dependence through the Clausius-Clapeyron relation. To compute radiative transfer we use PyRADS, a validated line-by-line column model (Koll & Cronin, 2019). We used the model in 1-D radiative-convective equilibrium, following Koll and Cronin (2018), in which a moist adiabat profile is assumed. We set the CO\textsubscript{2} concentration to 340 ppmv, the number of pressure levels to 30 (from 0.1 to 1,000 hPa), and consider a spectral range between 0.1 and 3,500 cm\textsuperscript{-1} at 0.01 cm\textsuperscript{-1} resolution. We only then need to specify the surface temperature and a vertically uniform relative humidity to compute OLR\textsubscript{cs} for the column. For more details of atmospheric structure and spectral databases, see “Materials and Methods” in Koll and Cronin (2018).

We calculate \( \lambda_{as} \) in the following way. We first compute OLR\textsub {cs} for some \( T_s \) and RH; we then perturb the surface temperature by an amount \( \Delta T_s \) and allow the moist-adiabatic atmosphere to respond while holding RH fixed; finally we calculate the perturbed OLR and take the finite difference between the two states. In summary:
where $\Delta T = 1$ K in our calculations. Note that the Planck, lapse rate, and water vapor feedbacks are included in $\lambda_{cs}$. The moist adiabat is not satisfied in the mid-latitudes, but we note that the lapse rate feedback is small when RH is fixed (Cess, 1975; Held & Shell, 2012; Jeevanjee et al., 2021; Zelinka et al., 2020). We exclude the RH-feedback associated with a change in RH with surface warming for simplicity and because its value in the global mean is $<0.1$ Wm$^{-2}$K$^{-1}$ (Held & Shell, 2012; Zelinka et al., 2020). We also assume the atmosphere responds like a moist adiabat for simplicity, although in reality the atmospheric temperature change is not always due to a local $T_s$ change (Mauritsen, 2016). We give some perspective on our feedback analysis in Section 3.1.

Our results are summarized in Figure 1a for surface temperatures between 230 and 320 K and relative humidities between 0% and 100%. We identify 275 K as the de facto boundary between a low temperature regime and a high temperature regime because each region exhibits distinct behaviors for $\lambda_{cs}$. Below 275 K, there is a very small RH-dependence — for values of RH between 20% and 80%, $\lambda_{cs}$ remains within 10% of

\begin{equation}
\lambda_{cs}(T_s, \text{RH}) \approx \frac{\text{OLR}_{cs}(T_s + \Delta T, \text{RH}) - \text{OLR}_{cs}(T_s, \text{RH})}{\Delta T},
\end{equation}
2.2 Wm$^{-2}$K$^{-1}$. Above 275 K, however, a significant RH-dependence emerges: the value of $\lambda_{2H}$ differs from 2.2 Wm$^{-2}$K$^{-1}$ by much more than 10% over the same range of humidity. We explicitly plot the RH-dependence of $\lambda_{2H}$ at 275 and 300K (Figure 1b) to highlight this difference in behavior. The majority of “Earth-like” values of ($I_{2H}$, RH) pairs fall within 10% of 2.2 Wm$^{-2}$K$^{-1}$ (indicated by the overlap between the boxed and stippled areas in Figure 1a). Thus, in response to Question 2, a $\lambda_{2H}$ value of 2.2 Wm$^{-2}$K$^{-1}$ will occur over much of the globe, and in particular will manifest as the constant slope of an OLR$_{2H}$ versus $T_s$ regression, as in Figure 1 of Koll and Cronin (2018). Nonetheless, Figure 1 shows that $\lambda_{2H}$ can still vary considerably at the higher temperatures of Earth’s tropics.

2.2. Importance of the H$_2$O Window

This section provides additional context about $\lambda_{2H}$ by focusing on the underlying radiation physics that controls the climate response.

Since $\lambda_{2H}$ is dominated by surface emission through the H$_2$O window (Koll & Cronin, 2018), the wavenumbers at which H$_2$O absorption is negligible for surface emission, we expect the window to play an important role in the RH-dependence of $\lambda_{2H}$ at high temperatures. To display the H$_2$O window, we plot the surface-to-space transmission $T_s$, which measures the portion of surface emission at wavenumber $\nu$ that escapes to space. At a surface temperature of 275 K (Figure 2a), the H$_2$O window remains open as the relative humidity is increased from 12.5% to 100%. At a surface temperature of 300 K (Figure 2b), however, the H$_2$O window closes rapidly as the relative humidity is increased from 12.5% to 100%. This is due to activation of H$_2$O continuum absorption (Koll & Cronin, 2018). Thus relative humidity variations are sufficient to close the H$_2$O window, but only at high temperatures.

Koll and Cronin (2018) emphasized the robustness of $\lambda_{2H} \approx 2.2$ Wm$^{-2}$K$^{-1}$ as arising from a balance between the closing of the H$_2$O window and the nonlinear $4\sigma T_s^4$ surface Planck feedback. However, at high temperatures this balance is not as robust, as evidenced by the decreasing width of the stippled area of the $\lambda_{2H}$ phase space, which denotes $\lambda_{2H} = 2.2 \pm 10$ Wm$^{-2}$K$^{-1}$, with increasing temperature (Figure 1a). The balance is not as robust because the H$_2$O window closes much faster with temperature at 100% RH than at 20% RH (Figure 2b), leading to a “Planck-dominated” response at low RH, and a “window-dominated” response at high RH (Figure 1c).

2.3. Tropical Variations in $\lambda_{2H}$

In Question 1 we asked whether the furnaces and radiator fins described in Pierrehumbert (1995) manifest as a contrast in the zonal-mean $\lambda_{2H}$ as opposed to the zonal-mean OLR$_{2H}$. To answer this question we will first compute time- and zonal-mean $I_{2H}$ and RH from ERA5 reanalysis (Hersbach et al., 2020) and use these values to read off zonal-mean $\lambda_{2H}$ from the phase space of Figure 1. We take hourly data sub-sampled every 6 h from January 1 to December 31, 1981 and compute annual averages. Following Zhang et al. (2020),
In Question 3, we asked whether clouds affect the meridional structure in zonal-mean longwave feedback in the tropics. Rather than explicitly compute cloud feedbacks, which is beyond the scope of this study, we try to estimate them by constructing a simple model for how clouds modify longwave emission. To validate the approach, we first estimate the all-sky OLR, \( \text{OLR}_{\text{cs}} \), from a few inputs: \( \text{OLR}_{\text{as}} \), the cloud top temperature \( T_c \), the OLR at 550 hPa, and RH in the tropics. We might expect a significant drop in \( \lambda_{cs} \) from the subtropics to the deep tropics by looking at representative values of \( T_c \) and RH in the \( \lambda_{cs} \)-phase space (Figure 1). Indeed, the meridional structure of zonal-mean \( \lambda_{cs} \) calculated as described above, shows that \( \lambda_{cs} \) varies from 2.6 to 1.6 Wm\(^{-2}\)K\(^{-1}\) in the tropics, a 38% drop (Figure 3a). Both the subtropical maxima and deep tropical minimum lie outside the 2.2 \( \pm 10\% \) Wm\(^{-2}\)K\(^{-1}\) range. We posit that these extrema of \( \lambda_{cs} \) should be considered the true “radiator fins” and “furnace,” respectively, of the tropics.

We can test whether the significant drop in \( \lambda_{cs} \) between the radiator fins and the furnace is due to the difference in humidity, as emphasized by Pierrehumbert (1995), or if the drop is due to the difference in temperature. If we look again at the phase space in Figure 1, we can take a path that goes from the subtropics to the deep tropics in two parts (the order does not matter): a first part with constant surface temperature, and a second part with constant relative humidity (see the dashed gray arrows). In this region of phase space, the doubling of relative humidity from 30% to 60% causes a much larger change in \( \lambda_{cs} \) than the increase in surface temperature from 295 to 300 K does.

Our answer to the first part of Question 1 is then: zonal-mean \( \lambda_{cs} \) exhibits local extrema, which may be usefully viewed as the “furnace” and “radiator fins” of the tropics. Furthermore, these extrema are indeed due to RH variations, consistent with Pierrehumbert (1995). \( \lambda_{cs} \) exhibits a local maximum in the subtropics because they are hot and dry enough for the feedback to exhibit a Planck-dominated response, and \( \lambda_{cs} \) exhibits a local minimum in the deep tropics because they are hot and moist enough for the feedback to exhibit a window-dominated response (Figure 3c).

To answer the second part of Question 1, that is, why radiator fins should not be regarded as a contrast in the zonal-mean \( \text{OLR}_{\text{cs}} \), we plot the annual- and zonal-mean \( \text{OLR}_{\text{cs}} \), and \( \text{OLR}_{\text{as}} \), from ERA5 reanalysis in gray in Figures 4c and 4d. Note the muted latitudinal variations of \( \text{OLR}_{\text{cs}} \) within the tropics (~10 Wm\(^{-2}\)) compared to variations in \( \text{OLR}_{\text{as}} \) throughout the rest of the globe (~100 Wm\(^{-2}\)). This muted latitudinal dependence within the tropics is inconsistent with the notion of radiator fins as significant subtropical maxima in \( \text{OLR}_{\text{cs}} \), which is why we focus on \( \lambda_{cs} \) instead. \( \text{OLR}_{\text{as}} \) does have more significant subtropical extrema, but these should not be interpreted as a furnace and radiator fins because the longwave warming effect of deep tropical clouds is balanced by their shortwave cooling effect (Hartmann & Berry, 2017; Pierrehumbert, 1995).

### 2.4. Incorporating the Effects of Cloudiness

In Question 3, we asked whether clouds affect the meridional structure in zonal-mean longwave feedback in the tropics. Rather than explicitly compute cloud feedbacks, which is beyond the scope of this study, we try to estimate them by constructing a simple model for how clouds modify longwave emission. To validate the approach, we first estimate the all-sky OLR, \( \text{OLR}_{\text{as}} \), from a few inputs: \( \text{OLR}_{\text{as}} \), the cloud top temperature \( T_c \), and the cloud top infrared OLR at 550 hPa, which is used to avoid misidentifying cloud tops with boundary layer cloudiness. \( T_c \) is the atmospheric temperature...
at which cloud fraction profile peaks. We smooth $T_{ct}$ with a Savitzky-Golay filter with a $10^\circ$ latitude width to account for sharp jumps in $T_{ct}$ arising from the limited vertical resolution. This method of identifying cloud tops is similar to Thompson et al. (2017). We show our methodology in the Supporting Information S1.

To estimate $OLR_{as}$, we first consider the effect of high clouds, which block longwave emission from lower levels and replace it with their own longwave emission from cloud tops. We assume high cloud emission acts like a black body and occurs high enough in the atmosphere that emission travels directly to space (Siebesma et al., 2020). As for low clouds, we grossly assume the low clouds emit at a temperature close enough to $sT_{f}$ that they only negligibly alter the outgoing radiation (Hartmann, 2015). Given these assumptions, we can now write down a simple expression for $OLR_{as}$:

$$O_{LRas} \approx \sigma T_{ct}^4 f + O_{LRcs}(1 - f).$$

This model is similar in some ways to the conceptual model created in Soden et al. (2008) to examine cross-field correlations between clear-sky and cloud feedbacks.

To get a sense of what the inputs to Equation 3 look like, we plot annual- and zonal-mean $Ef$, $T_{ct}$, $OLR_{cs}$, and $OLR_{as}$ from ERA5 reanalysis in gray in Figure 4. We test the approximate all-sky radiation from Equation 3 against $OLR_{as}$ directly output from ERA5 analysis, which includes cloud opacities and comprehensive radiative transfer in its calculation. We find that our model does an acceptable job in replicating the reanalysis (Figure 4d), although there is a slight underestimate within the tropics and a slight overestimate outside the tropics. Overall, the relative accuracy and physical transparency of our estimate gives us enough confidence in this model to proceed.

We now use Equation 3 to compute the longwave all-sky feedback, $\lambda_{as}$. We aim to assess the order of magnitude impact of clouds on our findings, so we first assume high cloud temperatures do not change appreciably with warming, consistent with the fixed anvil temperature (FAT) hypothesis (Hartmann & Larson, 2002; Zelinka & Hartmann, 2010). We assume a FAT as opposed to a proportionally higher anvil temperature (PHAT) to explore a well-defined limit of the high cloud altitude feedback. Although $Ef$ can change with warming (Bony et al., 2016; Saint-Lu et al., 2020), the high cloud area feedback is quite uncertain (Shewood et al., 2020; Wing et al., 2020), so for simplicity we assume $Ef$ is constant with warming. Differentiating Equation 3 with respect to $T_{ct}$ yields:

$$\lambda_{as} = \lambda_{cs}(1 - f).$$

This equation makes it conceptually clear how clouds modify $\lambda_{cs}$; the longwave feedback over clouds is 0. Since $f$ is positive definite (Figure 4), $\lambda_{as} \leq \lambda_{as}$, which is well demonstrated in the in the zonal mean in Figure 5a. The all-sky feedback looks like a simple translation downward of the clear-sky feedback, and there is still a significant (~50%) variation in $\lambda_{as}$ from the subtropics to the deep tropics. Our answer to Question 3 is then: Clouds have a destabilizing influence on the longwave feedback. However, the structure of all-sky feedback looks similar to clear-sky feedback, implying that the RH-dependence from clear-sky effects still dominates the meridional structure.
3. Discussion

Our work can be summarized as follows:

1. At high temperatures, variations in RH are sufficient to close the H₂O window, driving deviations in \( \lambda_{\text{cs}} \) from the typical value of 2.2 Wm⁻²K⁻¹ (Figure 1).

2. Furnaces and radiator fins can be interpreted as tropical extrema in zonal-mean \( \lambda_{\text{m}} \) as a consequence of the RH-dependence (Figure 3). They should not be interpreted as significant tropical extrema in zonal-mean OLR because tropical variations in OLR are small compared to global variations in OLR (Figure 4).

3. Cloud radiative effects can be estimated with a simple equation to reconstruct the all-sky OLR (Figure 5), which we then use to estimate the all-sky feedback. Clouds lower the feedback relative to clear skies, but the RH-dependence of the feedback remains significant (Figure 5).

3.1. Comparison to Other Work

We have demonstrated a reason for why a large contrast in \( \lambda_{\text{cs}} \) emerges in the tropics, but our results for the zonal-mean feedback cannot be directly compared to most other studies of regional feedback (e.g., Armour et al., 2013a and Feldl & Roe, 2013a, 2013b). Our clear-sky feedback \( \lambda_{\text{cs}} \) is not equal to the sum of the Planck, lapse rate, and water vapor feedbacks, because these feedbacks include “cloud climatological effects” (Yoshimori et al., 2020), that is, these feedbacks are calculated in the presence of clouds from the control simulation. Furthermore, our all-sky feedback \( \lambda_{\text{as}} \) is also not equal to this sum of conventional feedbacks, because those feedbacks fix the cloud pressure, whereas we fix the cloud temperature. These issues were discussed in detail by Yoshimori et al. (2020), and our \( \lambda_{\text{as}} \) should be comparable to their T-FRAT feedback in which the RH and cloud temperatures are fixed. Further work could explicitly explore such comparisons.

Our results can still be fruitfully compared to Zhang et al. (2020), who also analyzed the zonal-mean \( \lambda_{\text{cs}} \) in GCMs, but without the assumptions that column RH is fixed with warming and that the atmosphere follows a moist adiabat. We both find a drop in \( \lambda_{\text{cs}} \) of ~1 Wm⁻²K⁻¹ from the sub tropics to the deep tropics. However, Zhang et al. (2020) suggested that the drop in zonal mean \( \lambda_{\text{cs}} \) results from the RH-feedback due to local column RH increases with surface warming. Column RH is fixed in our study and yet we still get a significant tropical dip, although our \( \lambda_{\text{as}} \) is offset by a constant ~0.5 Wm⁻²K⁻¹ from their results. This comparison suggests that climatological RH causes the tropical variations in \( \lambda_{\text{cs}} \), whereas the RH-feedback and deviations from a moist adiabat uniformly lowers \( \lambda_{\text{as}} \). The importance of climatological RH is further supported by Bourdin et al. (2021), who also finds that climatological RH influences climate sensitivity, even if the vertical distribution of RH remains unchanged with warming.

Analyzing feedbacks locally or globally can give opposing impressions to the radiative response to warming. Local surface warming in the deep tropics yields a small, local OLR increase (as measured by \( \lambda_{\text{cs}} \) and \( \lambda_{\text{cs}} \)) but a large, global OLR increase (Dong et al., 2019). The discrepancy between the weak, local and the strong, global radiative response from warming the deep tropics arises from atmospheric temperature changes not associated with a local \( T_s \) change, that is, the remote warming of the free troposphere (Ceppi et al., 2017; Mauritsen, 2016). A local feedback analysis, by construction, cannot capture the effects of remote warming, which should be noted when comparing results across studies. The relative contributions of local surface warming versus non-local free-tropospheric warming to OLR change is relatively unexplored, so further study might alter our characterization of the tropical radiative response if one contribution can be shown to dominate over the other.

Our new understanding of the state dependence of \( \lambda_{\text{as}} \) gives context to previous results. For example, Bloch-Johnson et al. (2021) and Meraner et al. (2013) attributed an increase in equilibrium climate sensitivity to the decrease in \( \lambda_{\text{as}} \) with warming. We expect these variations in \( \lambda_{\text{as}} \) to be enhanced in climates that...
hotter than present-day Earth and conversely to be suppressed in climates cooler than present-day Earth. Our particular calculation of the $\lambda_s$ phase space assumed that the CO$_2$ concentration is fixed at 340 ppmv, which neglects the increasingly important role of CO$_2$ in stabilizing the climate at high CO$_2$ concentrations (Seeley & Jeevanjee, 2020). However, the strength in our approach of studying the joint dependence of $\lambda_s$ on $T_s$ and RH is its generality, for our approach cannot only be applied to our present-day climate, but to past climates like the Eocene, Pliocene, and Last Glacial Maximum, and to future climates predicted from different RCP scenarios.

Acknowledgments
This work was funded by CEMPS at the University of Exeter. The authors thank two anonymous reviewers for their constructive comments, which have greatly improved the manuscript. The authors also thank Penelope Maher, Stephen Thomson, Hugo Lambert, and Mitchell Koerner for their helpful conversations on this project.

Data Availability Statement
ERA5 data are available from https://doi.org/10.24381/cds.bd0915c6. Data from the PyRADS calculations are available from https://zenodo.org/record/5164050#.YqWwFlNKhZI.

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