

# Physical and observational constraints on the anvil cloud area feedback

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**Abstract** Changes in anvil clouds with warming remain a leading source of uncertainty in estimating the Earth’s climate sensitivity. Here, we develop a novel feedback analysis that decomposes changes in anvil clouds and creates testable hypotheses for refining their proposed uncertainty ranges with observations and theory. To carry out this storyline approach, we derive a simple but quantitative expression for the anvil area feedback, which is shown to depend on the present day, measurable cloud radiative effects and the fractional change in anvil area with warming. Satellite observations suggest an anvil cloud radiative effect of about  $\pm 1 \text{ Wm}^{-2}$ , which requires the fractional change in anvil area to be about  $\mp 50\% \text{ K}^{-1}$  to produce a feedback equal to the current best estimate of its lower bound. We use theory and observations to show that the change in anvil area is closer to about  $-4\% \text{ K}^{-1}$ . This constrains the area feedback and leads to our new estimate of  $0.02 \pm 0.07 \text{ Wm}^{-2}\text{K}^{-1}$ , which is many times weaker and more constrained than the overall anvil cloud feedback. In comparison, we show the anvil cloudy albedo feedback to be much less constrained, both theoretically and observationally, and thus to embody the uncertainty that poses an obstacle for bounding the Earth’s climate sensitivity.

1 **T**HE severity of global warming depends on how sensitive the  
2 Earth’s surface temperature is to increased  $\text{CO}_2$  (1, 2). De-  
3 termining the precise sensitivity hinges on constraining the cloud  
4 response to warming (3). While recent progress has been made in  
5 constraining low cloud feedbacks (4, 5), anvil cloud feedbacks re-  
6 main a large source of uncertainty (1, 2) despite decades of study  
7 (6–11). Anvil clouds pose a particular challenge because their ra-  
8 diative balance results from large, opposing radiative effects (12).  
9 Is this balance guaranteed? Or will warming tip the scales?

10 **Uncertainty in anvil cloud feedbacks** It has long been thought  
11 that anvil clouds might produce a strong negative climate feed-  
12 back: first by functioning as a shortwave *tropical thermostat* (6),  
13 however the observations that led to this conclusion are no longer  
14 considered evidence of such an effect (7, 13); then as an *infrared*  
15 *iris* (8). Criticisms of this latter study’s methodology soon fol-  
16 lowed (9, 14, 15), but they did not rule out the existence of a strong  
17 anvil cloud feedback.

18 Comprehensive assessments still consider changes in anvil  
19 clouds to be a leading source of uncertainty in estimates of cli-  
20 mate sensitivity (1, 2). These assessments refer to an uncertain  
21 *anvil cloud area* (or *amount*) *feedback*, but it is more accurate  
22 to regard it as the *altitude-corrected anvil cloud feedback* (here-  
23 after referred to as the *anvil cloud feedback*) since it results from  
24 changes in both area and optical depth, but not altitude changes  
25 which are considered in a separate feedback (16). The question of  
26 which individual feedback—area or optical depth—truly embod-  
27 ies the uncertainty remains unanswered. They must be separated  
28 to determine which is the main obstacle to constraining Earth’s  
29 climate sensitivity.

30 Anvil cloud area is controlled in part by unconstrained mi-  
31 crophysics (17–19), but also by robust thermodynamic principles  
32 (20, 21). These principles predict that anvils decrease in area with

warming because the static stability of the atmosphere increases  
(21), which is consistent with observed variability (22–24) and  
with most simulations (25). The resulting area feedback *might* be  
small because anvils are radiatively neutral (7, 26, 27). But how  
neutral must anvil clouds be for their area feedback to be insignif-  
icant? What if their radiative effect changes with warming? And  
what if when anvils shrink, more of the Earth is exposed to the  
radiative effects and feedbacks of underlying low clouds? Such  
questions have thus far limited our ability to constrain the anvil  
cloud area feedback.

Less is known about how cloud optical depth changes with  
warming, but it will manifest in optical properties such as the  
anvil’s cloudy albedo, the cloud reflectivity normalized by cloud  
fraction. Changes in cloudy albedo might produce an even stronger  
feedback than changes in area because anvils have a much stronger  
effect in the shortwave than in the net (26). But how much does  
cloudy albedo change with warming? And how much must it  
change to produce a substantial feedback? Such questions per-  
sist and so we cannot yet conclude whether this feedback is more  
uncertain than the other.

**Clearing the cloud of uncertainty** A physically-motivated  
decomposition that distinguishes the anvil area feedback from  
the anvil cloudy albedo feedback is needed. Since models must  
contend with representing unconstrained microphysics (17–19),  
we will primarily use observations. This rules out the use of purely  
model-based cloud feedback decompositions (28, 29). A cloud  
controlling factor analysis, an observational-based method mostly  
used for constraining low cloud feedbacks (30), requires further  
investigation before being suitable for constraining anvil cloud  
feedbacks with confidence. The connection between anvil clouds  
and the cloud controlling factors of the large-scale environment  
are not as well understood as in the case of low clouds.

Here, we derive a novel analytical cloud feedback decompo-  
sition based on the essential physics of cloud radiative effects.  
When combined with observations, this decomposition lets us

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68 identify, understand, and quantitatively constrain cloud feedbacks  
 69 in a physically transparent way. We adopt a *storyline approach*  
 70 (31), in which we examine the driving factors that control a cloud  
 71 feedback and judge the plausibility of these factors to produce a  
 72 particular feedback value by comparing to process evidence. This  
 73 approach shows which feedback is constrained and which is the  
 74 obstacle to constraining climate sensitivity.

## 75 Conceptualizing cloud radiative effects

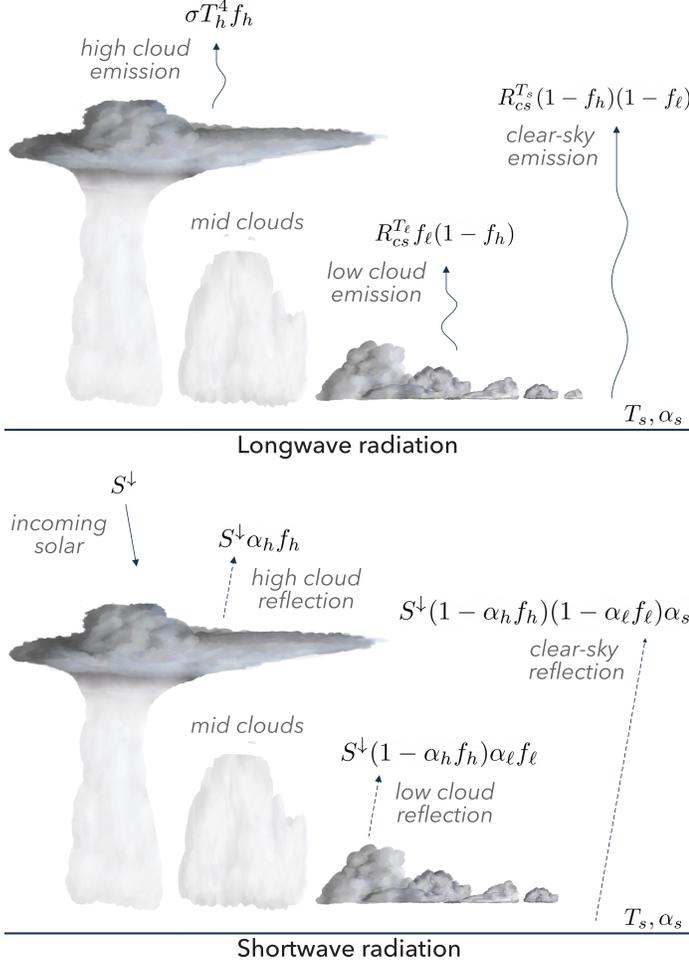


Figure 1: *Conceptualizing cloud radiative effects.* We idealize the vertical cloud profile into two distinct layers that represent anvil clouds and low clouds with random overlap. Equations indicate the domain-averaged contribution of high clouds, low clouds, and the surface to TOA energy balance. Their sum in the longwave and shortwave is given by Equation 14 and 16, respectively. See Table 1 for symbol meanings and values.

76 Clouds are complex, but for simplicity we divide them into  
 77 two types: high ( $h$ ) and low ( $\ell$ ). We subsume their properties into  
 78 a few bulk parameters that can be obtained from observations and  
 79 reanalysis (Table 1). These properties include their area fraction  
 80  $f_h, f_\ell$ , their emission temperature  $T_h, T_\ell$ , and their cloudy albedo  
 81  $\alpha_h, \alpha_\ell$  (which is cloud reflectivity normalized by cloud fraction).  
 82 Longwave emissivity will not be considered because most clouds  
 83 have an emissivity close to one (32). Clear-sky radiation can also  
 84 be distilled into a few parameters: the incoming solar radiation  
 85  $S^\downarrow$ , the surface albedo  $\alpha_s$ , and the outgoing longwave radiation  
 86 for a given surface temperature  $R_{cs}^{T_s}$ . Neglecting atmospheric

absorption will bias the surface and cloudy albedos to be higher  
 than they would otherwise be, but this simplification permits the  
 derivation of analytical expressions for cloud radiative effects from  
 high clouds and low clouds  $C_h, C_\ell$ ; cloud overlap effects  $m_{\ell h}$ ;  
 and the TOA energy balance  $N$ . See Figure 1 for an illustration  
 and Methods for the derivation. We will revisit some of these  
 assumptions when discussing cloudy albedo feedbacks

## Analytic feedbacks and the storyline approach

Feedbacks are computed by differentiating Earth's TOA energy  
 balance (Equation 16 minus Equation 14, see Methods) with respect  
 to the surface temperature  $T_s$  (33). To start, we have:

$$\lambda \equiv \frac{dN}{dT_s} = \frac{dN_{cs}}{dT_s} + \frac{dC}{dT_s}, \quad (1)$$

where  $N_{cs}$  is the clear-sky TOA energy balance and  $C = C_h + C_\ell + m_{\ell h}$  is the net cloud radiative effect from all clouds. Plugging in the analytical expressions for  $C$  (Equation 15 and 17, see Methods), we arrive at an equation for tropical climate feedbacks in terms of our bulk parameters:

$$\lambda = \lambda_0 + \sum_{i=h,\ell} \left( \lambda_i^{\text{area}} + \lambda_i^{\text{temp}} + \lambda_i^{\text{albedo}} \right), \quad (2)$$

where  $\lambda_0$  is the reference response assuming a fixed anvil temperature and fixed relative humidity (11, 34); and  $\lambda_i^{\text{area}}, \lambda_i^{\text{temp}}, \lambda_i^{\text{albedo}}$  are the feedbacks from changes in cloud area, cloud temperature, and cloudy albedo with warming. All feedbacks are described analytically. See Methods for the full derivation.

These analytic expressions form the basis of our storyline approach by transparently and quantitatively relating changes in cloud properties to their resulting radiative feedbacks. A more formal Bayesian framework of hypothesis testing (used in 1, 31 to constrain climate sensitivity with diverse lines of evidence) will not be necessary here because we primarily consider process evidence.

**The anvil cloud area feedback** Let us first focus on the high cloud area feedback,  $\lambda_h^{\text{area}}$ . After collecting all terms from Equation 1 that involve changes in anvil area  $df_h/dT_s$ , we arrive at a remarkably simple equation for the anvil cloud area feedback,

$$\lambda_h^{\text{area}} = \frac{d \ln f_h}{dT_s} (C_h + m_{\ell h}). \quad (3)$$

It depends on the *fractional* change in anvil area with warming  $d \ln f_h/dT_s$  and the sum of the *present day* anvil cloud radiative effect  $C_h$  and cloud overlap effect  $m_{\ell h}$ . The logarithmic derivative is used because fractional changes in cloud area are easier to interpret and bound than absolute changes—as we will soon see. And though we computed the change in cloud radiative effect with warming, our physically-based decomposition shows the area feedback does not depend on the change in radiative effect, but its present-day value. This means it can be quantified and used to constrain the feedback.

**The storyline approach in a nutshell** Equation 3 reveals that the smaller the climatological anvil cloud radiative effect, the larger the change in anvil area would have to be to produce a given feedback strength. Therefore, we can probe the plausibility of a particular strength by first quantifying the observed anvil cloud radiative effect; then calculating the change in anvil area required to produce such a feedback strength; and then comparing the required change in anvil area to the amount expected from theory, simulations, and observations. If the expected change in anvil area is much smaller than the required change, then that particular feedback strength can be ruled out.

## Climatology

Bounding the area feedback requires quantifying the tropically averaged anvil cloud radiative effect and cloud overlap effect ( $C_h + m_{\ell h}$ ). Since these quantities are not directly observed, they will be inferred from our simple model of cloud radiative effects.

We do this by inputting observations of cloud fraction from CALIPSO (35), clear-sky radiation from CERES (36), surface temperature from HadCRUT5 (37), and atmospheric temperature from ERA5 reanalysis (38) into our expression for the net cloud radiative effect (Equations 15 and 17), see Methods.  $f_h$  and  $f_\ell$  are identified as the maximum of the observed cloud fraction profile above and below 8 km, respectively. We then ensure goodness of fit between the inferred and the observed cloud radiative effects by adding a single scaling factor  $n$  to the anvil cloud fraction, which accounts for collapsing the anvil cloud fraction profile into a single level (Methods and Extended Data Figure 1). We treat  $n$  as a constant because spatio-temporal variations in the vertical profile of anvil clouds affect the optical depth and hence  $\alpha_h$  and  $\alpha_\ell$ , which already capture this variability as they are allowed to vary from year-to-year. In summary, we treat  $n$ ,  $\alpha_h$ , and  $\alpha_\ell$  as tuneable parameters to ensure consistency with observations at TOA (see Methods for further details).

We test our idealizations by comparing the observed net, shortwave, and longwave cloud radiative effects ( $C_{\text{obs}}$ ,  $C_{\text{obs}}^{\text{sw}}$ ,  $C_{\text{obs}}^{\text{lw}}$ ) with their counterparts from the simple model (Figure 2), which take the spatial fields of cloud fraction, temperature, albedo, and clear-sky radiation as inputs. Our model can reproduce the spatial patterns of longwave and shortwave cloud radiative effects, although there are small deviations throughout the tropics, such as an underestimate of  $C$  in the south east of China and an overestimate of  $C$  in the eastern Pacific, next to South America (Figure 2c). Given the overall close agreement, we consider our model fit for the task of evaluating the anvil cloud area feedback.

The climatological values of tropical quantities used in our calculations are summarized in Table 1 of Extended Data and the cloud properties of interest are plotted in Figure 3.  $f_h$  is maximum in the West Pacific Warm Pool and  $f_\ell$  is maximum along the East Pacific. Decomposing  $C$  into its contributions from different layers reveals that the net  $C$  is dominated by  $C_\ell$ . By comparison, the overlap effect  $m_{\ell h}$  is much smaller and varies less. The same is true for the high cloud radiative effect  $C_h$ , which exhibits a remarkable cancellation between its shortwave and longwave components not just in the warm pool (12, 24, 39), but across the tropics.

## Constraining the anvil cloud area feedback

With these more precise values in hand, we can constrain the tropical anvil cloud area feedback. To scale our estimate of  $\lambda_h^{\text{area}}$  to the global average, we multiply by the area ratio of the tropics and the globe, 1/2.

$$\langle \lambda_h^{\text{area}} \rangle = \frac{1}{2} \frac{d \ln f_h}{dT_s} (C_h + m_{\ell h}). \quad (4)$$

The plausible lower bound on  $\langle \lambda_h^{\text{area}} \rangle$  at present is  $-0.4 \text{ Wm}^{-2} \text{K}^{-1}$  which comes from assuming all the magnitude in the anvil cloud feedback in (1) is due to area changes alone. This lower bound allows the possibility of an overall negative cloud feedback, a necessary ingredient for a climate sensitivity below 1.5 K (31). Our inferred value tropical mean value of  $C_h + m_{\ell h} \approx -1.5 \text{ Wm}^{-2}$  implies that  $d \ln f_h / dT_s$  must be  $\approx 50\% \text{ K}^{-1}$  to achieve this feedback strength.

Following our storyline approach, we will assess how plausible these cloud changes are by comparing them to the changes expected from the stability iris hypothesis assuming a moist adiabat (21) and from observed interannual variability (22).

**Changes in anvil area with warming** The stability iris hypothesis (21) states that anvil cloud fraction  $f_h$  is proportional to detrainment from deep convection. Owing to mass conservation, this detrainment is equal to the clear-sky convergence,  $\partial_p \omega$ , where  $\omega$  is the subsidence vertical velocity [ $\text{hPa day}^{-1}$ ]. If we make the ansatz that  $\partial_p \omega$  is proportional to  $\omega$  at the level of detrainment ( $h$ ), then the fractional change in anvil area is equal to the fractional change in subsidence velocity at the anvil level:

$$\frac{d \ln f_h}{dT_s} = \frac{d \ln \omega_h}{dT_s}. \quad (5)$$

The subsidence velocity can be written as the quotient of the clear-sky radiative flux divergence in temperature coordinates ( $-\partial_T F$ ) and the difference between the actual and dry lapse rates (19):

$$\omega = \frac{-\partial_T F}{1/\Gamma - 1/\Gamma_d}. \quad (6)$$

Given that  $\partial_T F$  does not vary with surface temperature (40), if we further assume that  $\Gamma_h$ , the lapse rate at the anvil level, is moist adiabatic, then the change in cloud area can be computed with a few representative numbers. Assuming the surface warms from  $T_s = 298 \text{ K}$  to  $299 \text{ K}$  and the anvil cloud warms from  $T_h = 221 \text{ K}$  to anywhere between 221 and 221.4 K (a typical range of anvil warming, see 41 and references therein), then we expect that anvils change in area at about

$$\begin{aligned} \frac{d \ln f_h}{dT_s} &= - \frac{d \ln(1/\Gamma_h - 1/\Gamma_d)}{dT_s} \quad (\text{stability iris}) \\ &\approx -1 \text{ to } -4\% \text{ K}^{-1}, \end{aligned} \quad (7)$$

depending on the amount of anvil warming. Despite the numerous simplifications in our derivation, the result is similar to the mean and standard deviation of large-domain models in RCEMIP ( $-2 \pm 5\% \text{ K}^{-1}$  for cloud resolving models,  $-2 \pm 4\% \text{ K}^{-1}$  for all models, Table S5 of 41).

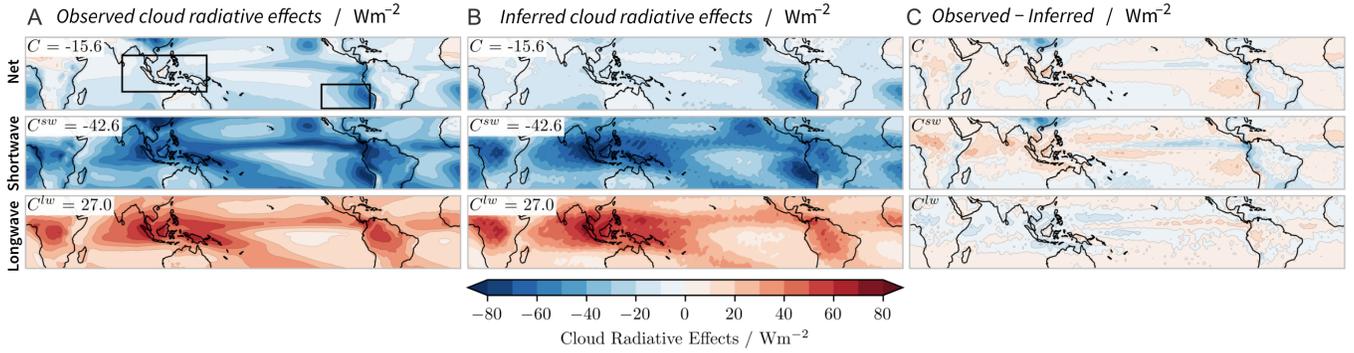


Figure 2: Observed net, shortwave, and longwave cloud radiative effects ( $C$ ,  $C^{sw}$ ,  $C^{lw}$ ) from CERES compared to their inferred counterparts. Tropical mean values are shown in the upper left of each panel. The West Pacific Warm Pool and East Pacific regions are boxed in a). The colorbar is the same for all plots.

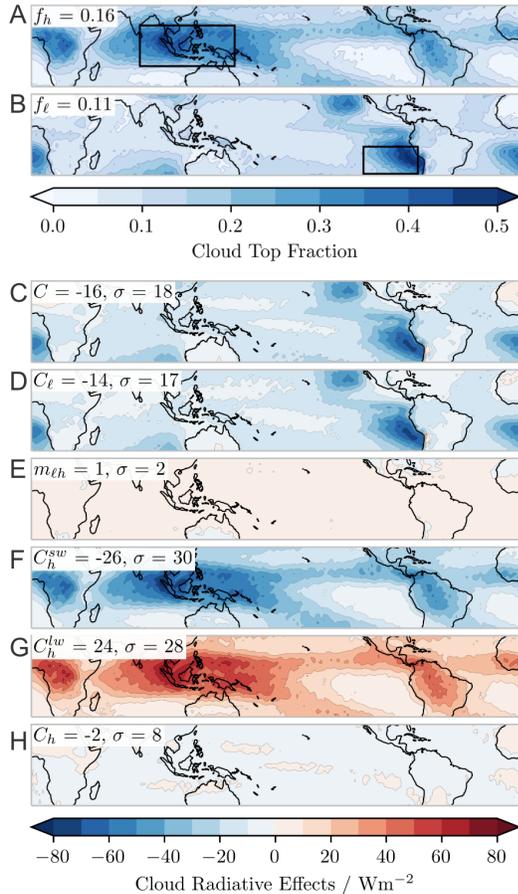


Figure 3: Climatological values of tropical quantities. Effective anvil cloud fraction (a) and low cloud fraction from CALIPSO (b). The West Pacific Warm Pool and East Pacific regions are boxed to indicate regions of maximum anvil and low cloud coverage, respectively. c-h) Inferred cloud radiative effects from Equations 18, 19, 21. Tropical mean values and standard deviations are shown in the upper left of each panel. Refer to Extended Data Figure 2 to see  $m_{lh}$  and  $C_h$  plotted with a finer color scale.

Now turning to ENSO-driven interannual variability, we compute annual averages of  $\ln f_h$  and  $T_s$  (the tropical mean surface temperature) from July to June, similar to (22). To avoid logarithmic divergences, we exclude grid cells with  $f_h = 0$ . We scatter annual averages of  $\ln f_h$  against  $T_s$  in Figure 4. The line of best fit for this relation gives

$$\frac{d \ln f_h}{dT_s} \approx -7 \text{ to } -11\% \text{ K}^{-1}. \quad (\text{interannual variability}) \quad (8)$$

This change is both larger than our simple estimate and from RCEMIP; it is also larger than the change of  $-5\% \text{ K}^{-1}$  inferred from interannual variability in AMIP runs with the IPSL, MPI, and NCAR models (see Figure S3 of 21). However, since all of these estimates of anvil cloud changes are much smaller than what is required to achieve our specified lower bound of  $\langle \lambda_h^{\text{area}} \rangle = -0.4 \text{ Wm}^{-2} \text{ K}^{-1}$ , the bounds of the area feedback can be refined.

**Best estimate of the area feedback** Care should be taken when determining the anvil cloud area change with warming on different timescales. Anvil area is better correlated with upper tropospheric stability than surface temperature (22, 23), and surface- and upper-tropospheric warming (and thus changes in stability) do not always go hand-in-hand on interannual timescales (42). This may alter the anvil area sensitivity to surface temperature inferred from variability. Indeed, anvil clouds are about half as sensitive for long term warming as compared to interannual variability in the IPSL general circulation model (23), the only model where such an analysis has been done. Furthermore, ENSO-driven interannual variability is not only associated with a change in surface temperature, but also a reorganization of deep convection from the West Pacific to the Central Pacific (43) which may further alter the inferred relationship between anvil area and surface temperature on different timescales.

Given the evidence from theory assuming a moist-adiabatic change in lapse rate (Equation 7), observations of interannual variability (Equation 8), and simulations (21, 23, 25), we estimate that the anvil cloud area changes at about

$$\frac{d \ln f_h}{dT_s} = -4 \pm 2\% \text{ K}^{-1}. \quad (\text{best estimate}) \quad (9)$$

We found  $C_h + m_{lh} = -1.5 \text{ Wm}^{-2}$ , but other observational studies have estimated  $-4 \text{ Wm}^{-2}$  (39),  $0.6 \text{ Wm}^{-2}$  (17), and  $2 \text{ Wm}^{-2}$  (44). This is probably due to methodological differences

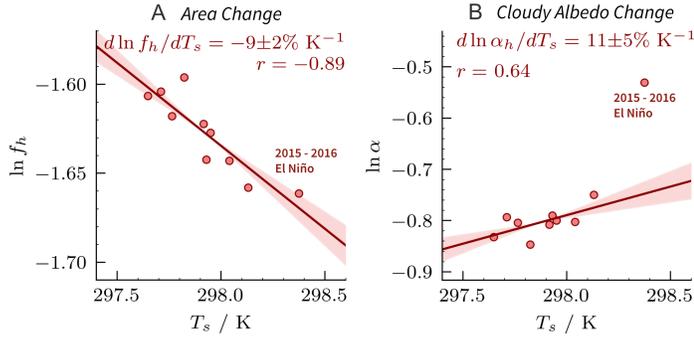


Figure 4: *Interannual changes in tropical mean anvil cloud area (a) and anvil cloudy albedo (b) as a function of surface temperature.* Each point represents one year from 2006–2016. In each subplot, the slope, correlation of the best fit line and its standard error are shown. Standard error in the slope due to limited sampling is indicated by shading. In (b), the regression is calculated excluding the 2015–2016 El Niño. See Extended Data Figure 4 for regression calculated including the El Niño and the regression calculated for low cloudy albedo.

and because anvil clouds have no precise definition. Furthermore, CERES TOA fluxes have their own small uncertainties (36) and considering mid-level clouds as distinct entities from low clouds adds an additional uncertainty of  $0.5 \text{ Wm}^{-2}$  (see Methods). Therefore, we estimate the anvil cloud radiative effect and cloud overlap effect to be,

$$C_h + m_{\ell h} = -1 \pm 3 \text{ Wm}^{-2}. \quad (\text{best estimate}) \quad (10)$$

Using these best estimates in Equation 4, we get our best estimate of the anvil area feedback to within one standard deviation:

$$\langle \lambda_h^{\text{area}} \rangle = 0.02 \pm 0.07 \text{ Wm}^{-2} \text{K}^{-1}. \quad (\text{best estimate}) \quad (11)$$

Overlap effects with low-level clouds are accounted for ( $m_{\ell h} = 0.5 \text{ Wm}^{-2}$ ): they dampen the anvil cloud area feedback by about 25%. Our estimate for the anvil cloud area feedback is positive but ten times smaller in magnitude and three times more constrained than the WCRP estimate of  $-0.2 \pm 0.2 \text{ Wm}^{-2} \text{K}^{-1}$  for the anvil cloud feedback (1). We deem that the area feedback is now well constrained because its uncertainty is comparable to other cloud feedbacks (1, 2). Our results provide a theoretical and observational basis for previously qualitative arguments regarding a small area feedback (7, 26, 27). What about the anvil cloudy albedo feedback?

## Uncertainty in anvil cloudy albedo feedback

Qualitative arguments and GCM experiments suggest a significant feedback could be produced without any change in anvil area (26, 29), but let us make that notion quantitative by considering the anvil cloudy albedo feedback,

$$\lambda_h^{\text{albedo}} = \frac{1}{2} \frac{d \ln \alpha_h}{d T_s} (C_h^{\text{sw}} + m_{\ell h}^{\text{sw}}). \quad (12)$$

It follows a similar form to the area feedback but depends on the fractional change in cloudy albedo with warming  $d \ln \alpha_h / d T_s$ , the shortwave anvil cloud radiative effect  $C_h^{\text{sw}}$ , and the shortwave cloud overlap effect  $m_{\ell h}$  (see Methods for derivation).

Given that  $C_h^{\text{sw}} + m_{\ell h}^{\text{sw}} \approx -25 \text{ Wm}^{-2}$  (Extended Data Table 1), producing a feedback of  $-0.2 \text{ Wm}^{-2} \text{K}^{-1}$  requires a fractional

change in cloudy albedo of only 1 to 2%  $\text{K}^{-1}$ . In contrast to anvil area, even a small change in the anvil’s cloudy albedo could produce a strong radiative response. How plausible is such a change?

**Changes in anvil cloudy albedo with warming** Computing the anvil cloudy albedos for each year (see Methods), we find that  $d \ln \alpha_h / d T_s \approx 6$  to  $16 \text{ \% K}^{-1}$  and significantly increases during the 2015–2016 El Niño (Figure 4b) for reasons that are unclear. Anvil height and temperature are not as sensitive to El Niño (Extended Data Figure 3); and changes in low cloudy albedo are more ambiguous than changes in anvil cloudy albedo (Extended Data Figure 4). Taken at face value, such a change in albedo implies  $\lambda_h^{\text{albedo}} \approx 1/2 \cdot 10\% \text{ K}^{-1} \cdot -25 \text{ Wm}^{-2} \sim \mathcal{O}(-1) \text{ Wm}^{-2} \text{K}^{-1}$ , a large negative feedback. This value should be interpreted carefully.

First, it is possible that our diagnosed values of cloudy albedo are biased by using an idealized two-layer model of cloud radiative effects that ignores clear-sky atmospheric absorption, assumes a spatially uniform cloudy albedo, and assumes a cloud emissivity of 1. We have shown an increase in anvil cloudy albedo with warming, whereas another observational study showed anvil cloud thinning with warming and thus a decrease in cloudy albedo (45). Yet another observational study showed ice water path, a proxy for optical depth, to be non-monotonic with sea surface temperatures (46).

Second, there is no guarantee that long term warming will follow interannual warming, as already discussed in the context of area changes.

Third, there are no quantitative theories to understand such a change. Anvil cloudy albedo might increase if anvils contain more condensate with warming, as could happen if precipitation efficiency remains constant or increases (21, 47), but the precise amount remains uncertain. It remains to be seen what happens to cloudy albedo in cloud-resolving and climate models, but those results would also warrant caution because of large intermodel spreads in climatology of cloud condensate and cloud radiative effects (41).

Fourth, if the anvil cloud optical depth is increasing, then its longwave emissivity  $\varepsilon_h$  will increase too. This will produce a countervailing positive longwave feedback,

$$\lambda_h^{\text{emissivity}} = \frac{1}{2} \frac{d \ln \varepsilon_h}{d T_s} (C_h^{\text{lw}} + m_{\ell h}^{\text{lw}}), \quad (13)$$

but with an uncertain magnitude (see Methods for further discussion).

The net result of these two competing components of the optical depth feedback are unclear, though they might be consistent with the negative anvil cloud feedback found in observations by Williams and Pierrehumbert (48) and which forms the basis of the feedback estimates in comprehensive assessments (1, 2). Given the lack of theory for understanding such changes, the conflicting observational evidence, and that there may be a countervailing longwave anvil emissivity feedback, we conclude that the magnitude and uncertainty of the anvil cloud feedback in these previous assessments is primarily embodied by optical depth changes and not by area changes. Changes in cloud-moisture coupling and convective aggregation are not considered here, but may further impact the TOA budget (49).

## Implications of uncertainty

We constrained the bounds on the anvil cloud area feedback by formulating a physical storyline for its prior bounds and then refuting that storyline with observations and theory. But in closing one door, we open another. The theoretical and observational uncertainty of anvil cloud optical depth changes precludes us from constraining its attendant feedbacks. While much attention has been devoted to understanding changes in anvil cloud area, it appears that focusing on optical depth changes will promise enhanced returns for constraining climate sensitivity.

A first step toward this goal might be to understand what happens to cloudy albedo during the 2015–2016 El Niño, perhaps with the aid of an improved theory of cloud condensate that provides mechanistic insight into the state variables that really matter for controlling changes in cloud optical properties. This would have manifold benefits for the other potential methods of constraining the anvil cloud optical depth feedback, including emergent constraints (50), model inter-comparisons (16, 25, 41), cloud-controlling factor analysis (3, 4, 30), process studies (5), and the use of climatological quantities to predict climate change (51), because confidence in these methods ultimately derives from understanding the physical relationships between environmental changes, cloud changes, and the TOA response.

Such a physically transparent approach has even broader implications. Communicating with the public about our confidence (or lack thereof) in clouds and climate change is hard. However, a physical theory of cloud feedbacks that can constrain, quantify, and interpret models and observations, like the one proposed here, could help clear the cloud of uncertainty.

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## Author Contributions

B.A.M and S.B. designed research; B.A.M. performed research. B.A.M, S.B, and J.L.D. analyzed data; and B.A.M wrote the paper.

## 571 Methods

572 **Data availability** CERES data were obtained from the NASA Langley  
 573 Research Center (<https://ceres.larc.nasa.gov/data/>). CALIPSO  
 574 /CLOUDSAT data were obtained from NASA Atmospheric Science Data  
 575 Center ([https://asdc.larc.nasa.gov/project/CALIPSO/CAL\\_LID-](https://asdc.larc.nasa.gov/project/CALIPSO/CAL_LID-L3_Cloud_Occurrence-Standard-V1-00_V1-00)  
 576 [L3\\_Cloud\\_Occurrence-Standard-V1-00\\_V1-00](https://asdc.larc.nasa.gov/project/CALIPSO/CAL_LID-L3_Cloud_Occurrence-Standard-V1-00_V1-00)). ERA5 reanalysis  
 577 data were obtained from the Copernicus Climate Change Service ([https://](https://cds.climate.copernicus.eu/)  
 578 [cds.climate.copernicus.eu/](https://cds.climate.copernicus.eu/)). HadCRUT5 data were obtained  
 579 from the Met Office Hadley Centre ([https://www.metoffice.gov.](https://www.metoffice.gov.uk/hadobs/hadcrut5/data/current/download.html)  
 580 [uk/hadobs/hadcrut5/data/current/download.html](https://www.metoffice.gov.uk/hadobs/hadcrut5/data/current/download.html)).

581 **Code availability** All scripts used to support the analysis of satellite  
 582 and reanalysis data will be made available in a Github repository upon  
 583 acceptance.

584 **Conceptualizing cloud radiative effects** We start with an idealized  
 585 model of cloud radiative effects at the top of the atmosphere (TOA). Al-  
 586 though tropical cloudiness is expected to be trimodal (52), for simplicity  
 587 we will consider a domain containing two cloud types: high clouds ( $h$ )  
 588 and low clouds ( $\ell$ ). (Many assessments of cloud feedbacks also use this  
 589 bi-modal decomposition (1).) Each type has an emission temperature  
 590  $T_h, T_\ell$ ; an optically thick cloud fraction  $f_h, f_\ell$ ; and an albedo  $\alpha_h, \alpha_\ell$   
 591 (Figure 1). Mid-level clouds will be considered in our error analysis.

592 The TOA energy balance is  $N = S - R$ , where  $S$  is the absorbed  
 593 shortwave radiation and  $R$  is the outgoing longwave radiation. The cloud  
 594 radiative effect  $C$  is the difference in  $N$  between all-sky and clear-sky  
 595 ( $cs$ ) conditions,  $C = N - N_{cs}$  (53).  $C$  can be decomposed into longwave  
 596 and shortwave components:  $C = C^{sw} + C^{lw}$ .

597 In the longwave component, clear-sky regions with a surface temper-  
 598 ature  $T_s$  will emit to space with an outgoing longwave radiation of  $R_{cs}^{T_s}$ ,  
 599 but a portion will be blocked by clouds. Longwave emissivity will not  
 600 be considered because most clouds have an emissivity close to one (32).  
 601 Assuming random overlap between high clouds and low clouds (54),  
 602 the domain-averaged clear-sky contribution is  $R_{cs}^{T_s}(1 - f_h)(1 - f_\ell)$ .  
 603 Low clouds are so close to the surface that we treat their emission to  
 604 space like clear-sky surface emission but at  $T_\ell$ . Their domain-averaged  
 605 contribution is  $R_{cs}^{T_\ell} f_\ell(1 - f_h)$ . Since  $R_{cs}^{T_s}$  is an approximately linear  
 606 function of temperature (55),  $R_{cs}^{T_\ell} \approx R_{cs}^{T_s} + \lambda_{cs}(T_s - T_\ell)$ , where  
 607  $\lambda_{cs} \equiv -dR_{cs}/dT_s \approx -2 \text{ Wm}^{-2}\text{K}^{-1}$  is a representative value for the  
 608 longwave clear sky feedback (34). We assume that high clouds are so  
 609 high that they emit directly to space (33) with a value  $\sigma T_h^4 f_h$ . Summing  
 610 these contributions, the domain-averaged outgoing longwave radiation is

$$R = R_{cs}^{T_s}(1 - f_h) + \sigma T_h^4 f_h + \lambda_{cs}(T_s - T_\ell)(1 - f_h)f_\ell, \quad (14)$$

611 and the longwave cloud radiative effect  $-(R - R_{cs})$  is

$$C^{lw} = R_{cs}^{T_s} f_h - \sigma T_h^4 f_h - \lambda_{cs}(T_s - T_\ell)(1 - f_h)f_\ell. \quad (15)$$

612 In the shortwave component, there is an incoming solar radiation  $S^\downarrow$ ,  
 613 and we assume that there is no absorption except at the surface. High  
 614 clouds reflect a portion  $\alpha_h f_h$  back to space. The transmitted radiation  
 615 then hits low clouds which reflect a portion  $\alpha_\ell f_\ell$  back to space (ignoring  
 616 secondary reflections with the anvils above). The transmitted radiation  
 617 then hits the surface which reflects a portion  $\alpha_s$  back out to space and  
 618 absorbs the rest. Summing these contributions, the domain-averaged  
 619 absorbed shortwave radiation at TOA is

$$S = S^\downarrow(1 - \alpha_h f_h)(1 - \alpha_\ell f_\ell)(1 - \alpha_s). \quad (16)$$

620 The TOA absorbed shortwave in clear-skies is  $S_{cs} = S^\downarrow(1 - \alpha_s)$ , so the  
 621 shortwave cloud radiative effect  $(S - S_{cs})$  is:

$$C^{sw} = S_{cs}(-\alpha_h f_h - \alpha_\ell f_\ell + \alpha_h \alpha_\ell f_h f_\ell). \quad (17)$$

622 It will prove helpful to separate the contribution of isolated high  
 623 clouds and isolated low clouds to the net cloud radiative  $C$ . Setting  
 624  $f_\ell = 0$  yields the isolated high cloud radiative effect:

$$C_h = (-S_{cs}\alpha_h + R_{cs}^{T_s} - \sigma T_h^4) f_h. \quad (18)$$

625 Setting  $f_h = 0$  yields the isolated low cloud radiative effect:

$$C_\ell = (-S_{cs}\alpha_\ell - \lambda_{cs}(T_s - T_\ell)) f_\ell. \quad (19)$$

626 The total cloud radiative effect  $C$  in terms of each cloud is:

$$C = C_h + C_\ell + m_{\ell h}, \quad (20)$$

627 where

$$m_{\ell h} = (S_{cs}\alpha_\ell\alpha_h + \lambda_{cs}(T_s - T_\ell)) f_\ell f_h, \quad (21)$$

628 represents the cloud overlap masking effect. Note that  $C_h \propto f_h, C_\ell \propto f_\ell$ ,  
 629 and  $m_{\ell h} \propto f_\ell f_h$ .

630 **Feedback decomposition** We will now derive various cloud feed-  
 631 backs from these equations and assume a fixed relative humidity. The  
 632 lapse rate feedback has been shown to be small when using this reference  
 633 response (56, 57), so it will be ignored here.

$$\begin{aligned} \lambda &\equiv \frac{dN}{dT_s} \\ &= \frac{S_{cs}}{dT_s} - \frac{dR_{cs}^{T_s}}{dT_s} + \frac{dC}{dT_s} \\ &= \lambda_{cs}(1 - f_h) \\ &\quad + (R_{cs}^{T_s} - \sigma T_h^4 + \lambda_{cs}(T_s - T_\ell)f_\ell - S_{cs}\alpha_h + S_{cs}\alpha_h\alpha_\ell f_\ell) \frac{df_h}{dT_s} \\ &\quad + (-\lambda_{cs}(T_s - T_\ell)(1 - f_h) - S_{cs}\alpha_\ell + S_{cs}\alpha_h f_h \alpha_\ell) \frac{df_\ell}{dT_s} \\ &\quad + -4\sigma T_h^3 f_h \frac{dT_h}{dT_s} \\ &\quad + -\lambda_{cs}(1 - f_h) f_\ell \frac{d(T_s - T_\ell)}{dT_s} \\ &\quad + (-S_{cs}f_h + S_{cs}f_h\alpha_\ell f_\ell) \frac{d\alpha_h}{dT_s} \\ &\quad + (-S_{cs}f_\ell + S_{cs}\alpha_h f_h f_\ell) \frac{d\alpha_\ell}{dT_s} \\ &\quad - S^\downarrow(1 - \alpha_h f_h)(1 - \alpha_\ell f_\ell) \frac{d\alpha_s}{dT_s} \\ &\quad - (T_s - T_\ell)(1 - f_h) f_\ell \frac{d\lambda_{cs}}{dT_s}. \end{aligned} \quad (22)$$

634 Recognizing that many of these terms can be rewritten as cloud  
 635 radiative effects, we get:

$$\begin{aligned}
\lambda = & \lambda_{cs}(1 - f_h) \\
& + \left( C_h + m_{\ell h} \right) \frac{d \ln f_h}{dT_s} \\
& + \left( C_\ell + m_{\ell h} \right) \frac{d \ln f_\ell}{dT_s} \\
& - 4\sigma T_h^3 f_h \frac{dT_h}{dT_s} \\
& - \lambda_{cs}(1 - f_h) f_\ell \frac{d(T_s - T_\ell)}{dT_s} \\
& + \left( C_h^{sw} + m_{\ell h}^{sw} \right) \frac{d \ln \alpha_h}{dT_s} \\
& + \left( C_\ell^{sw} + m_{\ell h}^{sw} \right) \frac{d \ln \alpha_\ell}{dT_s} \\
& + C_s \frac{d \ln \alpha_s}{dT_s},
\end{aligned} \tag{23}$$

where we have assumed that  $d\lambda_{cs}/dT_s$  is negligible, and  $C_s = -S^\downarrow(1 - \alpha_h f_h)(1 - \alpha_\ell)\alpha_s$  is the surface albedo radiative effect, which is equivalent to the ‘‘cryosphere radiative forcing’’ (58).

Now we name and then describe each term:

$$\lambda = \lambda_0 + \lambda_h^{\text{area}} + \lambda_\ell^{\text{area}} + \lambda_h^{\text{temp}} + \lambda_\ell^{\text{temp}} + \lambda_h^{\text{albedo}} + \lambda_\ell^{\text{albedo}} + \lambda_s^{\text{albedo}} \tag{24}$$

$\lambda_0$  is the anvil cloud-masked longwave clear-sky feedback. It is our null hypothesis for the climate response to warming because it assumes fixed relative humidity; fixed anvil temperature, area, and albedo; fixed low cloud temperature difference, area, and albedo; and fixed surface albedo.  $\lambda_h^{\text{area}}$  and  $\lambda_\ell^{\text{area}}$  are the feedbacks from a changing anvil cloud and low cloud area, respectively.  $\lambda_h^{\text{temp}}$  is the feedback from a changing anvil cloud temperature.  $\lambda_\ell^{\text{temp}}$  is the feedback from a changing temperature difference between low clouds and the surface.  $\lambda_h^{\text{albedo}}$ ,  $\lambda_\ell^{\text{albedo}}$ , and  $\lambda_s^{\text{albedo}}$  are the feedbacks from a changing albedo of anvil clouds, low clouds, and surface, respectively. We omit the surface albedo feedback from Equation 2 because we are interested in tropical climate.

For simplicity, we have assumed that cloud emissivities of high clouds and low clouds ( $\varepsilon_h, \varepsilon_\ell$ ) are equal to one (32). However, if we relax this assumption for completeness, one can show this leads to a high- and low-cloud emissivity feedback with the following form:

$$\begin{aligned}
\lambda_h^{\text{emissivity}} &= \left( C_h^{lw} + m_{\ell h}^{lw} \right) \frac{d \ln \varepsilon_h}{dT_s}, \\
\lambda_\ell^{\text{emissivity}} &= \left( C_\ell^{lw} + m_{\ell h}^{lw} \right) \frac{d \ln \varepsilon_\ell}{dT_s},
\end{aligned} \tag{25}$$

which closely resemble the form of the cloudy albedo feedback. Some of the other feedbacks will have small modifications, but they are unimportant here.

**Climatology** We combine monthly-mean satellite observations, surface temperature measurements, and reanalysis and re-grid all datasets onto a common  $2^\circ$  latitude  $\times$   $2.5^\circ$  longitude grid over the tropical belt ( $30^\circ\text{N}$ – $30^\circ\text{S}$ ) from June 2006 to December 2016. Although anvil clouds populate the globe (59), it is less clear how extratropical anvils change with warming. Most cloud feedback assessments only consider tropical anvil clouds, so we will follow this convention.

From the CALIPSO lidar satellite dataset (35), we obtain vertical profiles of cloud fraction for optical depths between  $0.3 \leq \tau \leq 5$ . This range excludes both deep convective cores and optically thin cirrus unconnected to deep convection (22). We then vertically smooth the native vertical 60 m resolution profiles with a 480 m running mean. For anvil detection, we consider ice cloud data above 8 km. For shallower clouds, we consider the sum of ice and liquid cloud fraction data below

8 km. The diagnosed cloud fractions are the absolute maximum of the profile in their respective domains, but if the identified maximum does not exceed a cutoff ( $f_{\text{cut}} = 0.03$ ), then that region is considered to be clear-sky ( $f = 0$ ). This algorithm is applied to every grid point and then tropically-averaged. Our approach thus far resembles (22), just extended to include low clouds.

To match the inferred cloud radiative effects with the observed, we consider an effective cloud fraction  $f_h = n \cdot \text{Max}(f(z))$  for high clouds, where  $n$  is a single tuned parameter to account for collapsing the high cloud profile into one level. This accounting is more important for high clouds, as their profile’s full width-half maximum is  $\approx 5$  km (Figure 1 of Extended Data), whereas low clouds are already localized with a full width-half maximum of  $\approx 1$  km (Figure 1 of Extended Data). While  $n$  could be more rigorously derived from detailed considerations of cloud overlap (54), we opt to determine  $n$  by fitting the predicted tropical- and time-averaged longwave cloud radiative effect  $C^{lw}$  to its observed counterpart  $C_{\text{obs}}^{lw}$  from CERES (see **Cloud fraction** section of Methods). Doing so yields a spatially and temporally constant value of  $n = 1.7$ . This value lies between that from assuming maximum overlap between each layer of the anvil cloud, which yields  $n = 1$  and random overlap, which yields  $n \approx 5$ .

The height of the diagnosed cloud fraction is then used to diagnose the cloud temperatures  $T_h, T_\ell$  at each space and time by selecting the corresponding atmospheric temperature in ERA5 reanalysis (38). We use the HadCRUT5 dataset (37) to diagnose the surface temperature  $T_s$ .

We use monthly mean TOA radiative fluxes, both clear-sky and all-sky, from the CERES satellite EBAF Ed4.1 product (36, 60). We diagnose the surface albedo  $\alpha_s$  as the ratio of upwelling clear-sky shortwave radiation  $S_{\text{cs}}^\uparrow$  to incoming shortwave radiation  $S^\downarrow$ . However, because shortwave absorption and scattering occurs in the real atmosphere, our surface albedo is more accurately characterized as the planetary clear-sky albedo (61). We diagnose the cloud albedos by assuming that they are constant in space, and by fitting the predicted tropical- and time-averaged shortwave cloud radiative effect  $C^{sw}$  to its observed counterpart  $C_{\text{obs}}^{sw}$  from CERES. With two unknowns we must provide two constraints. We do this by splitting the tropics into two distinct dynamical regimes based on threshold of 500 hPa midtropospheric velocity  $\omega_{500} = 25 \text{ hPa day}^{-1}$  obtained from monthly ERA5 reanalysis data. These regions are treated as independent so that they provide two constraints. The regime-averaged shortwave radiative effect is then fitted to its observed counterpart by using the `fsolve` function from the `scipy.optimize` python module. (The precise threshold of  $25 \text{ hPa day}^{-1}$  was chosen because it resulted in the smallest root mean square error between  $C^{sw}$  and  $C_{\text{obs}}^{sw}$ .)

**Cloud fraction** We use the CALIPSO Lidar Satellite CAL\_LID\_L3\_Cloud\_Occurrence-Standard-V1-00 data product (62), the same dataset used in (22). While the high cloud fraction could simply be diagnosed as the maximum cloud fraction of the profile (i.e.  $f_h = \text{Max}(f(z))$ ), the calculated longwave cloud radiative effect  $C^{lw}$  will not match with observations. To rectify this, we will consider using a single tuning parameter,  $n$ . That is, we have an effective cloud fraction  $f_h = n \cdot \text{Max}(f(z))$  which accounts for representing a cloud profile with a single level.

We first demand that  $n$  be constant with space and time to ensure that areal changes (changes in  $f_h$ ) are not artificially convolved with vertical changes which relate to optical depth and albedo ( $\alpha$ ). This decision projects the spatio-temporal variability in the vertical extent of anvils more onto  $\alpha$  than  $f_h$ .

We then fit the predicted tropically- and temporally-averaged longwave radiative effect  $C^{lw}$  to its observed counterpart  $C_{\text{obs}}^{lw}$  from CERES. Given these constraints, and the inputs to Equation 15,  $n$  can be solved for as

$$n = \frac{\langle C_{\text{obs}}^{\text{lw}} + \lambda_{\text{cs}}(T_s - T_\ell)f_\ell \rangle}{\langle R_{\text{cs}} \max(f(z)) - \sigma T_h^4 \max(f(z)) + \lambda_{\text{cs}}(T_s - T_\ell)f_\ell \max(f(z)) \rangle}, \quad (26)$$

733 where  $\langle \cdot \rangle$  denotes a tropical- and temporal-average.

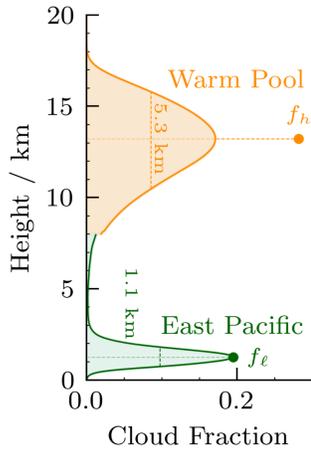
734 **Uncertainty analysis for area feedback** Uncertainty in our esti-  
 735 mates of  $d \ln f_h / dT_s$  and  $C_h + m_{\ell h}$  translate to uncertainty in  $\lambda_h^{\text{area}}$ . As  
 736 stated in the main text, we estimate  $d \ln f_h / dT_s = -4 \pm 2 \%$   $\text{K}^{-1}$ . For  
 737 the anvil cloud radiative effect, we found  $C_h + m_{\ell h} = -1.5 \text{ Wm}^{-2}$ .  
 738 However, other observational studies have found it to be  $-4 \text{ Wm}^{-2}$  (39),  
 739  $0.6 \text{ Wm}^{-2}$  (17), and  $2 \text{ Wm}^{-2}$  (44). This is probably due to methodolog-  
 740 ical differences and the fact that anvil clouds have no precise definition.  
 741 Furthermore, CERES TOA fluxes monthly fluxes have a stated uncer-  
 742 tainty of  $2.5 \text{ Wm}^{-2}$  (36).

743 Another source of error comes from neglecting mid-level clouds, a  
 744 fairly common cloud type (52), as their own identities. Let's assume  
 745 that emission from mid level congestus clouds ( $c$ ) experience a clear-sky  
 746 greenhouse effect. By symmetry with low clouds, they should contribute  
 747 an additional cloud overlap masking term that appears in our expression  
 748 for  $\lambda_{\text{area}}$ :  $m_{ch} = (S_{\text{cs}} \alpha_c \alpha_h + \lambda_{\text{cs}}(T_s - T_c))f_c f_h$ . Assuming that  $f_c =$   
 749  $0.1$ ,  $f_h = 0.17$ ,  $\alpha_c = \alpha_h = 0.45$ ,  $T_c = 250 \text{ K}$ ,  $T_s = 298 \text{ K}$ ,  $S_{\text{cs}} = 347$   
 750  $\text{Wm}^{-2}$ ,  $\lambda_{\text{cs}} = -2 \text{ Wm}^{-1} \text{K}^{-1}$  yields  $m_{ch} \approx -0.5 \text{ Wm}^{-2}$ .

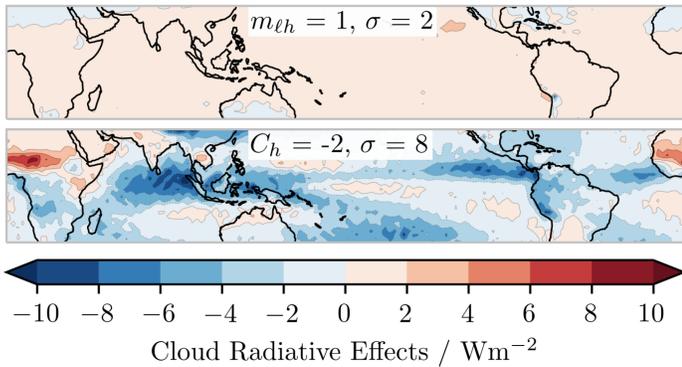
751 We therefore estimate  $C_h + m_{\ell h} = -1 \pm 3 \text{ Wm}^{-2}$ . This results in  
 752 our best estimate of the anvil cloud area feedback:

$$\begin{aligned} \langle \lambda_h^{\text{area}} \rangle &= 1/2 \cdot (-4 \pm 2 \%) \text{K}^{-1} \cdot (-1 \pm 3 \text{ Wm}^{-2}) \\ &= 0.02 \pm 0.07 \text{ Wm}^{-2} \text{K}^{-1}. \end{aligned} \quad (27)$$

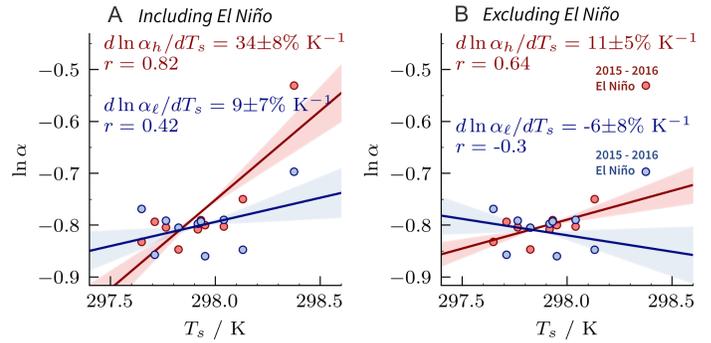
753 **Further uses of our framework** Our feedback expressions might  
 754 also provide a quick, quantitative, and physically transparent way to  
 755 interpret how model biases influence feedbacks. For instance, if members  
 756 of a GCM ensemble simulate  $C_h$  between  $\pm 10 \text{ Wm}^{-2}$ , but they all  
 757 simulate the same  $d \ln f_h / dT_s = -4 \%$   $\text{K}^{-1}$ , then their area feedbacks  
 758 will range between  $\mp 0.2 \text{ Wm}^{-2} \text{K}^{-1}$ . If all ensemble members simulate  
 759  $C_h = 1 \text{ Wm}^{-2}$ , but simulate  $d \ln f_h / dT_s = \pm 5 \%$   $\text{K}^{-1}$ , then their  
 760 area feedbacks will range between  $\pm 0.03 \text{ Wm}^{-2} \text{K}^{-1}$ . This quantitative  
 761 yet clear diagnostic could provide testable hypothesis that advance our  
 762 understanding and development of models.



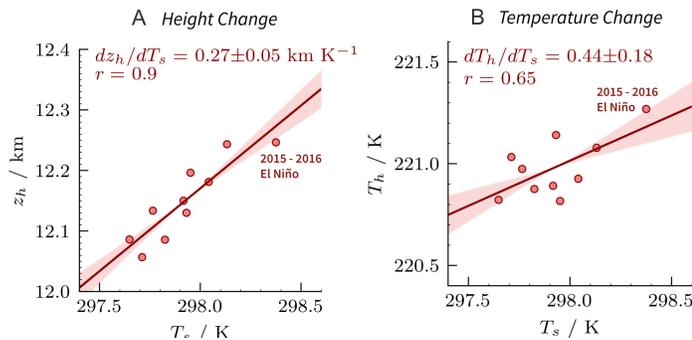
Extended Data Figure 1: *Illustration of effective cloud fraction.* The high cloud fraction profile in the Warm Pool and low cloud fraction profile in the East Pacific are from CALIPSO. The full width-half maximum and effective cloud fraction of each profile are shown. The high cloud and low cloud profiles are clipped below 8 km and above 4 km, respectively, in accordance with our detection method.



Extended Data Figure 2: *Climatological values of tropical quantities.* Top) Inferred cloud overlap effect from Equation 21. Bottom) Inferred anvil cloud radiative effect from Equation 18. Tropical mean values and standard deviations are shown in the upper middle of each panel. Refer to Figure 3 to see  $m_{lh}$  and  $C_h$  and other quantities plotted with a broader color scale.



Extended Data Figure 4: *Interannual changes in tropical mean anvil cloudy albedo (red) and low cloudy albedo (blue).* (a) The line of best fit is calculated with the 2015–2016 El Niño included. (b) The line of best of fit is calculated without the El Niño.



Extended Data Figure 3: *Interannual changes in tropical mean anvil cloud height (a) and temperature (b).* In each subplot, the slope, correlation for the best fit line and its standard error are shown. Standard error in the slope due to limited sampling is indicated by shading.

Extended Data Table 1: *Climatological values of tropical quantities (30°S – 30°N) used in this study.* All radiative quantities are evaluated at the top of atmosphere.  $C_{\text{obs}}^{\text{lw}}$  and  $C_{\text{obs}}^{\text{sw}}$  refer to the observed longwave and shortwave cloud radiative effects from CERES. See Climatology section for details.

Quantity	Description	Tropical mean value	Derivation
$f_h$	Anvil cloud area fraction	0.16	CALIPSO
$f_\ell$	Low cloud area fraction	0.11	CALIPSO
$T_h$	Anvil temperature	221 K	ERA5
$T_\ell$	Low cloud temperature	287 K	ERA5
$T_s$	Surface temperature	298 K	HadCRUT5
$\alpha_s$	Planetary surface albedo	0.13	CERES
$S^\downarrow$	Incoming shortwave radiation	398 $\text{Wm}^{-2}$	CERES
$S_{\text{cs}}$	Clear-sky absorbed shortwave	347 $\text{Wm}^{-2}$	CERES
$R_{\text{cs}}$	Clear-sky outgoing longwave	287 $\text{Wm}^{-2}$	CERES
$n$	Effective cloud fraction scaling	1.7	Fitted from $C_{\text{obs}}^{\text{lw}}$
$\alpha_h$	Anvil albedo	0.45	Fitted from $C_{\text{obs}}^{\text{sw}}$
$\alpha_\ell$	Low cloud albedo	0.45	Fitted from $C_{\text{obs}}^{\text{sw}}$
$C$	Net cloud radiative effect	-15.6 $\text{Wm}^{-2}$	Inferred
$C^{\text{sw}}$	Shortwave cloud radiative effect	-42.6 $\text{Wm}^{-2}$	Inferred
$C^{\text{lw}}$	Longwave cloud radiative effect	27.0 $\text{Wm}^{-2}$	Inferred
$C_h$	Anvil cloud radiative effect	-2.0 $\text{Wm}^{-2}$	Inferred
$C_h^{\text{sw}}$	Shortwave anvil cloud radiative effect	-26.1 $\text{Wm}^{-2}$	Inferred
$C_h^{\text{lw}}$	Longwave anvil cloud radiative effect	24.1 $\text{Wm}^{-2}$	Inferred
$C_\ell$	Low cloud radiative effect	-14.3 $\text{Wm}^{-2}$	Inferred
$C_\ell^{\text{sw}}$	Shortwave low cloud radiative effect	-17.3 $\text{Wm}^{-2}$	Inferred
$C_\ell^{\text{lw}}$	Longwave low cloud radiative effect	3.1 $\text{Wm}^{-2}$	Inferred
$m_{\ell h}$	Cloud overlap effect	0.6 $\text{Wm}^{-2}$	Inferred
$m_{\ell h}^{\text{sw}}$	Shortwave cloud overlap effect	1.1 $\text{Wm}^{-2}$	Inferred
$m_{\ell h}^{\text{lw}}$	Longwave cloud overlap effect	-0.5 $\text{Wm}^{-2}$	Inferred