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# Clear-sky control of the response of anvils to CO<sub>2</sub>, surface warming and volcanic eruptions

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## ABSTRACT

Anvil clouds produced by deep convection cover extensive areas of the tropics, and their response to external perturbations matters for the Earth's climate sensitivity. It has been suggested that variations in the height<sup>1-3</sup> and spatial extent<sup>3</sup> of these clouds can be understood from basic physical arguments related to the conservation of mass and energy in the clear-sky areas of the tropics. Based on satellite observations, meteorological reanalyses and climate model simulations, we show that these arguments can be used to interpret the response of anvil cloud fraction to a range of perturbations in the current climate and under climate change. This includes the response to interannual and long-term surface temperature changes, the response to the direct effect of carbon dioxide, and the decrease of anvil cloud fraction after explosive volcanic eruptions. Therefore, the control of tropical anvils by clear-sky radiative cooling and static stability in the upper troposphere can explain a large diversity of the responses of anvil cloud fraction to natural and anthropogenic perturbations. These findings should also be considered when assessing the impacts of geo-engineering techniques.

The response of high clouds to global warming is one of the largest sources of uncertainty for future projections<sup>2;4-7</sup>. Much of the tropical high-level cloudiness is composed of anvil clouds, which form at the top of convective systems at around 12-13 km<sup>8</sup>. Theoretical arguments<sup>1;2</sup> suggest that their altitude follows the upper-tropospheric peak of mass convergence in clear-sky regions, which rises near-isothermally in response to surface warming.

In clear-sky regions, subsidence ( $\omega_r$ , in Pa/s, positive downwards) is driven by the clear-sky radiative cooling rate ( $Q_r$ , defined such that cooling is positive):

$$\omega_r = \frac{Q_r}{S} \text{ with } S = \frac{T}{p} \frac{R}{c_p} - \frac{\partial T}{\partial p}, \quad (1)$$

where  $p$  is the air pressure,  $S$  is the atmospheric static stability,  $T$  is the temperature,  $R$  is the gas constant and  $c_p$  is the isobaric specific heat of dry air. The strong pressure gradient of  $Q_r$  in the upper-troposphere implies a maximum  $D_r$  in the horizontal convergence, itself directly dependent on the pressure gradient of subsidence:

$$D_r = \max \left( \frac{\partial \omega_r}{\partial p} \right). \quad (2)$$

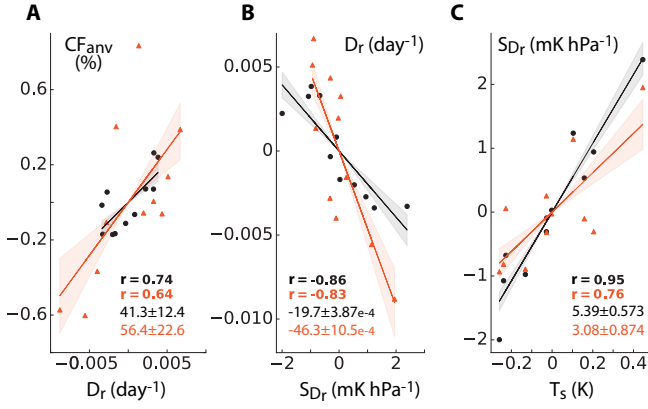
Owing to mass continuity, a maximum  $D_r$  of horizontal mass convergence in clear-sky regions is associated with a maximum of horizontal mass divergence in convective regions, and therefore with the fraction and altitude of anvil clouds. This argument has been invoked to predict a decrease of anvil cloud fraction with temperature, in the framework of the stability-Iris mechanism<sup>3</sup>. This mechanism is based on thermodynamic arguments and supported by radiative-convective equilibrium simulations from General Circulation

Models (GCMs) and cloud-resolving models<sup>3;9;10</sup> as well as observations<sup>8;11</sup>.

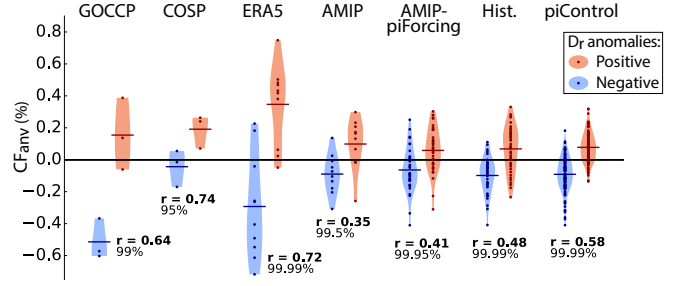
When the surface warms, anvils and  $D_r$  rise nearly isothermally, and find themselves in a more stable atmosphere because of the dependency of  $S$  on atmospheric pressure. Due to enhanced  $S$ , the clear-sky pressure gradient of subsidence is reduced, reducing  $D_r$ , which by mass conservation leads to reduced anvil cloud fraction.

The question arises as to whether similar arguments can explain the behavior of anvil cloud fraction under a range of natural and anthropogenic perturbations, associated or not with changes in surface temperature. Indeed, the stability-Iris mechanism is based on the relationship between anvil cloud fraction and the maximum  $D_r$  in clear-sky radiative convergence: surface temperature constitutes only one factor that can affect  $D_r$ . Perturbations independent of surface temperature variations also have the potential to cause changes in anvil cloud fraction, through variations in the atmospheric radiative cooling and temperature profile. This includes the purely radiative direct effect of CO<sub>2</sub><sup>12;13</sup> and other radiative perturbations in the stratosphere and upper troposphere induced by ozone changes<sup>14</sup> or volcanic eruptions.

Here we test this hypothesis by using a combination of satellite observations (GOCCP<sup>15</sup>), meteorological reanalyses (ERA5<sup>16</sup>) and different simulations of the current and future climate from a GCM (IPSL-CM6A-LR<sup>17</sup>). First, we show that this GCM reproduces the stability-Iris effect observed on interannual timescales. Then we examine the different responses of anvil clouds to increased CO<sub>2</sub> concentrations: the long-term response which is mediated by surface temperature changes, and the response to the direct effect of CO<sub>2</sub> which is not mediated by surface temperature changes<sup>12;18</sup>. Finally, we explore the response of anvil cloud fraction to ex-



**Fig. 1.** Relationships between (A)  $CF_{anv}$  and  $D_r$ , (B)  $D_r$  and  $S_{Dr}$ , (C)  $S_{Dr}$  and  $T_s$ , where  $CF_{anv}$  is the anvil cloud fraction derived from GOCCP observations (orange) and from the IPSL AMIP experiment using the COSP simulator (black) (see Methods),  $D_r$  and  $S_{Dr}$  are the upper-tropospheric clear-sky maximum of horizontal mass convergence and the static stability at the level of this maximum, derived from the ERA5 reanalysis (orange) interpolated on the IPSL spatial grid, and from IPSL AMIP (black), and  $T_s$  is the surface temperature derived from IPSL AMIP, where SSTs are derived from observations (orange and black). Lines show linear regression with shadings showing the regression error. Bold numbers are Pearson correlation coefficients (all p-values are below 0.05). Non-bold numbers are slopes. All quantities are tropical averages (30N-30S) July-to-June annual anomalies relative to the period 2006-2017 (see Methods).



**Fig. 2.** Violin plots showing  $CF_{anv}$  interannual anomalies for years of positive (red) and negative (blue)  $D_r$  anomalies, in: GOCCP observations (2006-2017), the IPSL AMIP COSP simulator (2006-2017), ERA5 (1980-2017), IPSL AMIP (1980-2017), AMIP-piForcing (1870-2017), historical (1850-2014) and piControl (455 years). Positive and negative  $D_r$  anomalies are taken from the last and first quartiles of the  $D_r$  anomalies, respectively. Horizontal bars give the distribution mean. Bold number are Pearson correlation coefficients between  $CF_{anv}$  and  $D_r$  anomalies including all years (all p-values are below 0.05). Percentage numbers are confidence intervals of the Welch t-test, testing that the mean  $CF_{anv}$  anomaly of the red distribution is strictly greater than that of the blue distribution. All anomalies are tropical averages (30N-30S) July-to-June annual anomalies relative to an 11-year running window.

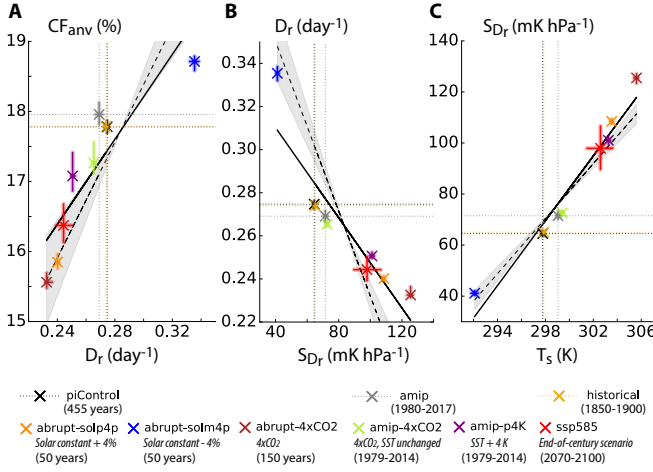
the relationship between  $CF_{anv}$  and  $D_r$  is very robust in the IPSL model as it is found in even longer simulations, and in a wide range of configurations, including atmosphere-only and ocean-atmosphere coupled simulations, with and without anthropogenic forcing (see Methods).

## Anvil cloud response under climate change

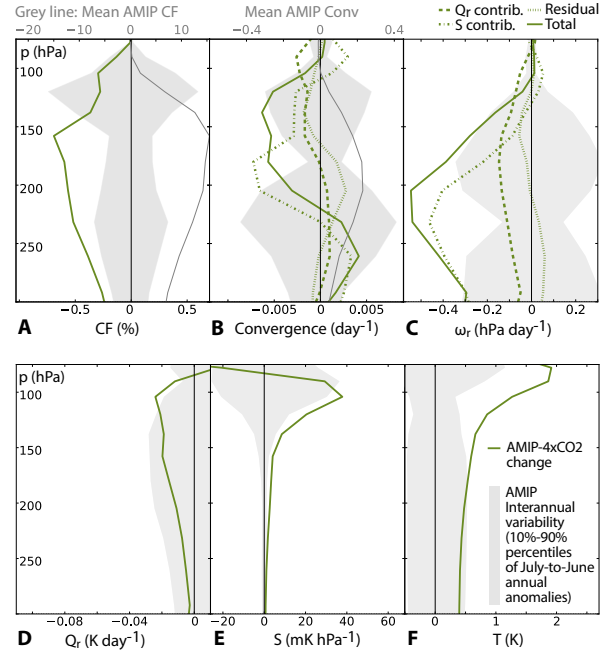
Since the IPSL model reproduces the relationship between  $CF_{anv}$  and  $D_r$  observed in the present-day climate, we use it to investigate the response of anvil cloud fraction under climate change.

As the climate warms in response to the increase of greenhouse gases in the atmosphere,  $CF_{anv}$  and  $D_r$  are both reduced, as shown in Fig. 3A by the changes in the abrupt-4xCO<sub>2</sub> experiment and in the future scenario SSP585. This long-term response mediated by surface temperature changes is not specific to the long-term effects of CO<sub>2</sub> forcing (Fig. 3A): it is also found when SSTs are forced to increase by 4 K, as shown by the changes in AMIP-p4K, or when surface temperatures are affected by long-term solar perturbations, as shown by the changes in abrupt-solp4p and abrupt-solm4p (where the solar constant is increased and decreased by 4%, respectively). With increasing  $T_s$ ,  $S_{Dr}$  is increased (Fig. 3C) by the upward shift of the isotherms towards lower pressures, reducing  $D_r$  (Fig. 3B) and hence  $CF_{anv}$  (Fig. 3A) following the stability-Iris mechanism (as shown in Fig. S1 by AMIP-p4K atmospheric profiles compared to AMIP).

The climate response to an increase in CO<sub>2</sub> concentration is partly due to surface warming and partly due to the direct effect of CO<sub>2</sub>. In the latter case, the large-scale control of anvils by  $D_r$  is also found, as shown in Fig. 3A by the reductions of both  $CF_{anv}$  and  $D_r$  in a simulation where CO<sub>2</sub> is quadrupled but SSTs remain unchanged (AMIP-4xCO<sub>2</sub>). These changes are due to the effect of increased CO<sub>2</sub> on clear-sky radiative



**Fig. 3.** Scatterplots showing long-term annual means of (A)  $CF_{anv}$  against  $D_r$ , (B)  $D_r$  against  $S_{Dr}$ , (C)  $S_{Dr}$  against  $T_s$ , between nine different climates simulated by the IPSL model (see Methods). To compare SSP585 with its reference historical climate, the last 30 years of SSP585 and the first 50 years of the historical experiment are shown. Error bars show the range of July-to-June yearly means. Black lines show linear regressions. To compare with interannual relationships, the slopes of Fig 1 for AMIP are shown here by the dashed lines with their errors as shadings. All quantities are tropical averages (30N-30S).



**Fig. 4.** Annual mean changes in (A) cloud fraction, (B) clear-sky horizontal mass convergence  $\partial\omega_r/\partial p$ , (C) clear-sky subsidence, (D) clear-sky radiative cooling (with cooling positive), (E) static stability and (F) temperature, in AMIP-4xCO2 (see Methods) compared to AMIP. Changes in convergence and subsidence are decomposed into the estimated parts due to the  $Q_r$  and  $S$  changes (dashed lines; see Methods). All vertical profiles are tropical averages (30N-30S).

cooling and atmospheric temperature (Fig. 4). The weakening of the clear-sky radiative cooling ( $Q_r$ , Fig. 4D) reduces the radiatively-driven subsidence ( $\omega_r$ , Fig. 4C and Eq. 1), reducing the clear-sky horizontal mass convergence in the upper-troposphere ( $\partial\omega_r/\partial p$ , Fig. 4B), hence reducing  $D_r$  (Fig. 3A and Eq. 2). At the same time, the weakening of  $Q_r$  increases the upper-tropospheric temperature ( $T$ , Fig. 4F) and hence the stability ( $S$ , Fig. 4E), further reducing the clear-sky horizontal mass convergence and hence  $D_r$ . The latter increase in  $S$  contributes the most to the reduction in clear-sky subsidence and convergence (Fig. 4C and B). As a result, the upper-tropospheric cloud fraction is reduced ( $CF$ , Fig. 4A). Unlike when the surface temperature increases, in which case the reduction in  $D_r$  is caused by the increase in stability associated with the rise of the isotherms towards lower pressures, here the altitude of  $D_r$  and hence of anvils does not rise (Fig. 4A and B, Fig. S2). Consistent reductions in  $CF_{anv}$  are found in other IPSL experiments (Fig. S2).

This reduction of  $D_r$  due to the direct effect of  $CO_2$  is gradually canceled when the surface temperature increases (Fig. S1). Indeed, in this case, the increase in  $T$  increases  $Q_r$  by increasing emission of infrared radiation to space, canceling its initial decrease in the upper-troposphere and hence canceling its effect on  $D_r$ . Therefore, the reduction of  $CF_{anv}$  due to the direct effect of  $CO_2$  is not expected to add up linearly with the reduction of  $CF_{anv}$  due to surface warming, but is instead expected to be canceled by the latter and its associated rise of  $D_r$ .

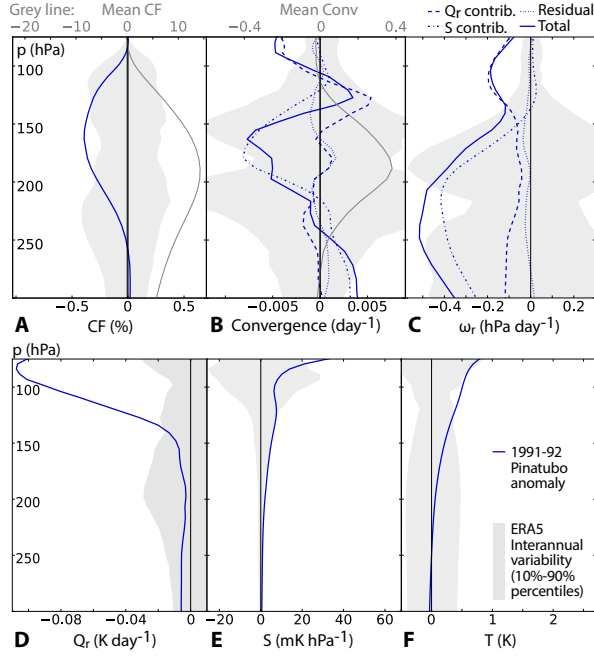
When considering all the different perturbations discussed above, we find that the sensitivity of  $CF_{anv}$  to  $D_r$  appears consistent with its sensitivity to interannual variations in  $D_r$  (Fig. 3A). In contrast, the sensitivity of  $D_r$  to  $S_{Dr}$  appears weaker than at interannual timescales (Fig. 3B). This is be-

cause  $D_r$  and  $S_{Dr}$  depend on several quantities ( $Q_r$ ,  $T$  and  $p$ ) whose variation is specific to the type of perturbation applied to the climate system (Fig. S3). On the other hand, the relationship between  $D_r$  and  $CF_{anv}$  which is central to the stability-Iris effect, appears to be more universal across different timescales and perturbations.

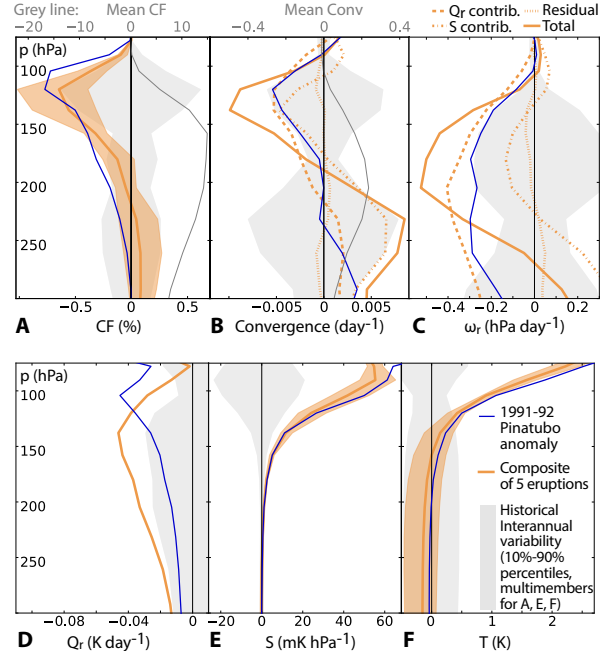
## Anvil cloud response to volcanic eruptions

While the abrupt change in  $CO_2$  concentration is idealised, explosive volcanic eruptions are abrupt changes that are observable. These events lead to aerosol increase in the stratosphere and upper-troposphere<sup>19</sup> that modifies the temperature and radiative cooling and thus has the potential to affect  $D_r$  and  $CF_{anv}$ . We test this hypothesis with the ERA5 reanalysis and the IPSL model, which were both able to reproduce the observed  $CF_{anv}$ - $D_r$  relationship on the 2006-2017 period (Fig. 1). There was no explosive volcanic eruption during the 11 years (2006-2017) of high-resolution space-borne lidar observations. Therefore we analyze the ERA5 reanalysis, whose cloud radiative effect has been shown to be very consistent with observations<sup>20</sup>.

Vertical profiles from ERA5 show that anvil cloud fraction is reduced during the year following the large volcanic eruption of Mount Pinatubo in 1991 (Fig. 5A). The lower stratosphere and upper troposphere warm by up to 1 K (Fig. 5F) due to the absorption of solar and infrared radiation by stratospheric aerosols. As for the direct effect of  $CO_2$ , this warming is associated with a reduced  $Q_r$  and increased  $S$  (Fig. 5D



**Fig. 5.** As Fig. 4 but during the year post-Pinatubo (blue) in ERA5, with all profiles except mean ones (grey line) being July-to-June annual anomalies relative to an 11-year running window (over 1980-2017). All profiles are tropical averages (30N-30S).



**Fig. 6.** As Fig. 5 but during the year post-Pinatubo in IPSL AMIP (blue), and for a composite average of annual anomalies for 6 different years that are subsequent to 5 major volcanic eruptions (1883-84, 1884-85, 1902-03, 1912-13, 1963-64, 1991-92) in IPSL historical (orange). When available (A, E and F) the historical composite profile is averaged over 11 historical members and orange shading shows the inter-members spread.

and  $E$ ), which both act to reduce  $\omega_r$  (Fig. 5C) and hence the clear-sky convergence in the upper-troposphere (Fig. 5B), in association with reduced  $CF$  (Fig. 5A). Near the anvils level (150-200 hPa), the reduction in  $Q_r$  is small; it is thus the increase in  $S$  which contributes the most to the reduction in clear-sky subsidence and convergence (Fig. 5C and B).

The IPSL AMIP experiment qualitatively exhibits the same behavior as ERA5 for the year following the Pinatubo eruption (Fig. 6). The historical experiment (1850-2014) includes five explosive volcanic eruptions, associated with the world-wide strongest and longest anomalies on stratospheric aerosols optical depth in the historical forcing<sup>21</sup>: Krakatoa (1883, very strong aerosol forcing persisting until 1885), Santa Maria (1902), Novarupta (1912), Mount Agung (1963-64) and Mount Pinatubo (1991). Fig. 6 shows a composite of atmospheric anomalies associated with these five events, averaged over 11 realizations of the historical experiment<sup>22</sup> (except for  $Q_r$ ,  $\omega_r$  and  $\partial\omega_r/\partial p$  which are shown for only one historical realization). The results are consistent with the Pinatubo eruption from ERA5, although here the reduction in  $Q_r$  strongly contributes to the reduction in clear-sky subsidence and convergence, even more so than the increase in  $S$  (Fig. 6C and B).

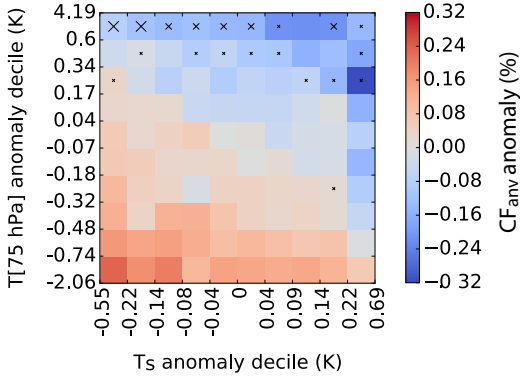
The reduction in radiative convergence and anvil cloud fraction during major volcanic eruptions is also found in the absence of anthropogenic forcing and SST variations (Fig. S4). Besides, there is no microphysical coupling between aerosols and ice clouds in the IPSL model<sup>23</sup>, nor in the ERA5 reanalysis<sup>16</sup>, which assures that aerosols can not directly affect anvil clouds formation. The reduction in radiative convergence and anvil cloud fraction during major volcanic eruptions is thus a large-scale atmospheric response to

volcanic eruptions, without any mediation by SST variations, anthropogenic forcing or microphysical effects.

This large-scale response of anvil clouds to volcanic eruptions is also consistent with Boucher et al. (2017)<sup>24</sup>, who found in a previous version of the IPSL atmospheric model that stratospheric sulfate aerosol injections — as considered for geo-engineering — reduce the tropical high cloud cover. They interpreted this reduction as caused by the warming and stabilization of the lower stratosphere and upper troposphere, which can be elucidated by our mechanism. Since sulfate aerosol injections have been shown to warm the lower stratosphere in several GCMs<sup>25;26</sup>, and even to reduce the optical depth of thin ice clouds<sup>25</sup>, these results are unlikely to be specific to the IPSL model. The IPSL model has also been shown to be consistent with other GCMs in terms of predicting the climate response to the Pinatubo eruption<sup>27</sup>.

Large volcanic eruptions are known to cool the surface, which would tend to increase  $CF_{anv}$  if  $D_r$  were primarily perturbed by surface cooling. The increase in  $CF_{anv}$  does not occur because the  $D_r$  perturbation is dominated instead by the effect of volcanic eruptions on lower-stratospheric temperatures. This is confirmed by Fig. 7, which shows that for a given stratospheric temperature anomaly,  $CF_{anv}$  decreases as  $T_S$  increases (consistent with the stability-Iris effect), and for a given  $T_S$  anomaly,  $CF_{anv}$  decreases when the stratosphere gets warmer (consistent with the effect of volcanic eruptions). The volcanic eruptions of the 11 historical realizations are associated with the most strongly positive stratospheric temperature anomalies. As a result, they remain largely associ-





**Fig. 7.** Diagram showing the  $CF_{anv}$  anomaly for each decile of surface temperature anomaly ( $T_S$ ) and lower-stratospheric temperature anomaly ( $T[75 \text{ hPa}]$ ), in 11 concatenated members of the IPSL historical experiment. Anomalies associated with the largest volcanic eruptions (1883-84, 1884-85, 1902-03, 1912-13, 1963-64, 1991-92) are indicated with crosses. Crosses can be of small, medium or large size: small if they represent only one anomaly among the 66 anomalies associated with volcanic eruption (6 years times 11 members), medium if they represent between 2 and 9 anomalies, and large if they represent 10 anomalies or more. All quantities are tropical averages (30N-30S) July-to-June annual anomalies relative to an 11-year running window.

ated with negative  $CF_{anv}$  anomalies regardless of the surface temperature. In these cases,  $D_r$  and  $CF_{anv}$  anomalies are dominated by stratospheric temperatures and  $T_S$  anomalies only play a secondary role, even when they are extremely cold (most occurrences of volcanic eruptions project in the two coldest deciles of  $T_S$ ). Volcanic eruptions as well as other occurrences of an extremely hot or cold stratosphere, can thus overcome the effect of  $T_S$  variations on  $CF_{anv}$ .

## Summary and Discussion

The idea that the relationship between anvil cloud fraction  $CF_{anv}$  and clear-sky radiative convergence  $D_r$  could induce  $CF_{anv}$  changes in response to  $T_S$  changes has been suggested on the basis of physical arguments and idealized RCE simulations<sup>3</sup>, and then confirmed by observations in an 11-year period<sup>8</sup>. Reanalyses and climate simulations both show that this mechanism operates on much longer timescales, both in response to natural  $T_S$  variations and externally-induced changes in  $T_S$ , such as in response to  $CO_2$  or solar constant perturbations. This shows the robustness of the stability-Iris mechanism.

However,  $D_r$  is not only affected by variations in  $T_S$ . We show that the direct effect of  $CO_2$  also perturbs  $D_r$  through its impact on  $Q_r$  and  $S$  and, consistently, that it is associated with changes in  $CF_{anv}$ .

We further show that explosive volcanic eruptions also perturb  $D_r$  and hence  $CF_{anv}$ , in a similar manner as  $CO_2$ . This study thus proposes a new mechanism through which stratospheric aerosols can affect the high-level cloudiness, which is not rooted in the microphysical aerosol-cloud interactions: here, stratospheric aerosols affect anvil cloudiness remotely through their impact on  $Q_r$  and  $S$ . According to this mechanism, the tropical high cloud cover is expected to shrink in

response to stratospheric sulfate aerosol injections, as actually found by Boucher et al. (2017)<sup>24</sup>. It should therefore be considered when investigating the potential impacts of geo-engineering techniques.

Our study shows that the response of  $CF_{anv}$  to a large diversity of natural and anthropogenic perturbations can be interpreted from the same basic mechanism rooted in the conservation of mass and energy. A recent assessment pointed out the large uncertainty associated with the anvil cloud fraction feedback for climate sensitivity<sup>7</sup>. It is thus encouraging that at least one robust control of  $CF_{anv}$  by a basic physical mechanism has been identified. Further studies will be necessary to understand the radiative implications of  $CF_{anv}$  changes on the Earth's radiative budget.

## Methods

**Calculation of  $D_r$  and  $S_{D_r}$ .** Following Eq. 2,  $D_r$  is defined (in each location and month) as the maximum value, between 600 hPa and 10 hPa, of the pressure gradient  $\partial\omega_r/\partial p$ . At the height of this maximum, the value of  $S$  (computed from Eq. 1) is referred to as  $S_{D_r}$ .

**Tropical anvil cloud fraction** We define anvil clouds following the detection method of Saint-Lu et al.<sup>8</sup>: we consider the vertical profile of monthly cloud fraction above 8 km, at each grid cell in the tropics (30°N-30°S), and we select the local maximum that is the closest to its centroid. This local maximum defines the anvils cloud fraction  $CF_{anv}$ .

**July-to-June annual anomalies** Monthly data are averaged from July of each calendar year to June of the next calendar year, giving July-to-June annual data, to detect interannual variations associated with El Niño-Southern Oscillation<sup>8</sup>. The annual-mean data over an 11-year period is then subtracted to this July-to-June annual data, to give July-to-June annual anomalies.

## Comparison between IPSL-CM6A-LR and observations

To best assess the realism of the anvil clouds behavior against observations, we use a simulation in which SSTs, land use, greenhouse gases, aerosols and ozone are all consistent with observations, which is the AMIP experiment<sup>28</sup>. We also use the COSP satellite simulator<sup>29</sup> to ensure that the definition of the cloud fraction is consistent between the model and the CALIPSO observational product GOCCP<sup>15</sup>. Note that one of the realizations of the AMIP experiment run with IPSL-CM6A-LR is extended to December 2017, covering the 11 years of space-borne lidar observations, allowing us to robustly compare model and observations.

**GCM experiments** Experiments used in this paper are all part of CMIP6 (Coupled Model Intercomparison Project phase 6).

- DECK experiments<sup>28</sup>: AMIP (atmospheric simulation with anthropogenic forcing and prescribed observed SSTs, taken over the period 1980-2017 - see above), piControl and historical (ocean-atmosphere coupling with pre-industrial 1850 forcing applied over 455 years and historical forcing applied over 1850-2014, respectively), abrupt-4xCO2 ( $CO_2$  multiplied by 4 compared to piControl at the beginning of the simulation and then held constant with time; using the last 150 years of the simulation).
- CFMIP experiments<sup>30</sup>: AMIP-piForcing (same as AMIP but with constant pre-industrial atmospheric forcing and run over 1870-2017), abrupt-solp4p and abrupt-solm4p (solar constant

increased and decreased by 4% compared to piControl, respectively; 50 year-long), AMIP-p4K (same as AMIP but SSTs are subject to a uniform warming of 4 K, over 1979-2014), AMIP-4xCO<sub>2</sub> (same as AMIP but CO<sub>2</sub> is quadrupled, over 1979-2014).

- Climate change scenarios<sup>31</sup>: SSP585 (2015-2100).

*Contributions of clear-sky radiative cooling and static stability changes to clear-sky subsidence and convergence changes* The change in clear-sky radiative subsidence  $\Delta\omega_r$ , can be decomposed from Eq. 1 as:

$$\Delta\omega_r = \frac{\Delta Q_r}{\bar{S}} - \bar{Q}_r \frac{\Delta S}{\bar{S}^2} \quad (3)$$

where overbars denote the time-mean for the reference state from which the change or anomaly is computed.

The change in clear-sky radiative convergence  $\Delta(\partial\omega_r/\partial p)$  can be approximated as  $\partial(\Delta\omega_r)/\partial p$  and decomposed as:

$$\frac{\partial(\Delta\omega_r)}{\partial p} = \frac{1}{\bar{S}} \left( \frac{\partial(\Delta Q_r)}{\partial p} - \frac{\Delta Q_r}{\bar{S}} \frac{\partial \bar{S}}{\partial p} \right) \quad (4)$$

$$+ \frac{1}{\bar{S}} \left( -\frac{\partial \bar{Q}_r}{\partial p} \frac{\Delta S}{\bar{S}} - \frac{\bar{Q}_r}{\bar{S}} \frac{\partial(\Delta S)}{\partial p} + 2\bar{Q}_r \frac{\partial \bar{S}}{\partial p} \frac{\Delta S}{\bar{S}^2} \right) \quad (5)$$

In both equations, the first and second terms on the right hand side give estimates of the contributions of  $Q_r$  and  $S$ , respectively.

## Authors contributions

M.S. conducted the study, analyzed the data and wrote the paper. All authors designed the research, interpreted the results and edited the manuscript.

## Data availability statement

GOCCP 3D cloud fraction are available at [https://climserv.ipsl.polytechnique.fr/cfmip-obs/Calipso\\_goccp.html](https://climserv.ipsl.polytechnique.fr/cfmip-obs/Calipso_goccp.html). ERA5 monthly means of daily means (MODA) on single levels are available on the Climate Data Store (CDS) online catalogue, at <https://cds.climate.copernicus.eu/#!/search?text=ERA5>, except for clear-sky mean temperature tendency due to long-wave and short-wave radiation (mttlwracs, mttswracs) used to compute  $Q_r$ , which can be downloaded in a raw format from the CDS API client following instruction at <https://confluence.ecmwf.int/display/CKB/How+to+download+ERA5#HowtodownloadERA5-4-DownloadERA5familydatathroughtheCDSAPI>. CMIP6 data are available through the Earth System Grid Federation (ESGF)<sup>32</sup>, see <https://pcmdi.llnl.gov/CMIP6/Guide/dataUsers.html#3-accessing-model-output>

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