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# North Pacific surface warming pattern shaped by atmospheric responses and radiative feedbacks

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## 25    **Abstract**

26        Greenhouse warming features spatial variations in sea surface temperature (SST)  
27    that modulate circulation, precipitation, and radiation. While South Pacific SST pattern  
28    has recently been explained by Southern-Hemisphere feedbacks, North Pacific  
29    warming remains poorly understood. Here we decompose coupled SST pattern  
30    formation into the effects of atmospheric responses to a spatially uniform SST increase  
31    (SUSI) and the direct CO<sub>2</sub> effect, radiative feedbacks, and air-sea interactions. Our heat  
32    budget shows significant mid-latitude warming driven by a stability response due to  
33    warm-air advection, reinforced by the surface radiative effects of melting sea ice and  
34    rising atmospheric moisture. Moderately enhanced subtropical warming reflects  
35    anomalous latent heating in SUSI: weakened surface winds associated with reduced  
36    subtropical anticyclone, and increased relative humidity due to greater evaporation  
37    minus precipitation. Thus, unlike in the South Pacific, the North Pacific warming  
38    pattern involves surprisingly little coupled air-sea interaction, being shaped instead by  
39    forced atmospheric responses and clear-sky radiative feedbacks.

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## Main text

During greenhouse warming, sea surface temperature (SST) change is spatially non-uniform<sup>1,2</sup>. Defined as the deviation of SST increase from its tropical mean, the SST patterns ( $T^*$ ) exert powerful controls on atmospheric circulation<sup>3,4</sup>, hydroclimate<sup>5,6</sup> and climate feedbacks<sup>7-9</sup>. This links global-scale radiative forcing to the regional climate changes experienced by societies and ecosystems. Understanding the formation process is critical to providing effective observational constraints<sup>10</sup>. The tropical Pacific exhibits two prominent  $T^*$  modes: an eastern equatorial peak, or El Niño-like warming<sup>11</sup>, and a subtropical south-to-north gradient<sup>2,12</sup>. The former has been studied extensively<sup>1,11,13-16</sup>, yet the processes determining the latter remain poorly constrained despite its robustness and importance<sup>3-12</sup>.

The cross-equatorial wind-evaporation-SST (WES) feedback (Methods) has long been proposed as the key mechanism for this subtropical gradient<sup>17-20</sup>. However, the relative SST cooling in the South Pacific, associated with strengthened trade winds, was recently explained by a local, single-hemisphere, inclined WES (SHIW) feedback<sup>12</sup> rather than a cross-equatorial process. Intermodel correlations of the subtropical SST and wind speed between the hemispheres are insignificant, undermining the idea of a single WES causing opposite anomalies across the equator. The North Pacific warming should therefore be examined as a physically distinct, single-hemisphere process. It remains unclear whether this is affected by large-scale forced responses, local radiative feedbacks, or coupled air-sea interactions.

To elucidate the North Pacific SST pattern formation, we decompose<sup>12</sup> changes under the 1% CO<sub>2</sub> increase per year (1pctCO2) scenario into those in spatially uniform SST increase (SUSI), direct radiative effect (RAD), and the residual  $T^*$ -driven component. As the first two components artificially prevent SST patterns from occurring, using atmosphere-only models, they represent the atmospheric responses to uniform SST warming and greenhouse forcing, before any  $T^*$  driving. Here, radiative feedbacks can exist to some extent, e.g., the large-scale water vapor plus lapse rate

feedback due to tropospheric warming that increases with height<sup>8</sup>. However, they primarily occur in the atmosphere and we still categorize them as the responses. The final effect includes local radiative feedbacks (shortwave and longwave, clear and cloudy skies) and coupled air-sea interactions (surface turbulent fluxes and oceanic heat transport). Combining this with a mixed-layer heat budget establishes a novel response-feedback-interaction framework for examining SST warming ([Methods](#)).

[Fig. 1](#) exhibits a hook-shaped SST pattern (green curved arrow) around the North Pacific Subtropical Anticyclone<sup>21,22</sup> (NPSA). We distinguish two warming regimes: the strong mid-latitude string (blue trapezoid) and the moderate subtropical tip (orange parallelogram). The string will be shown to result from surface stability increase by anomalous warm surface air temperature (SAT) advection and radiative effect of winter sea-ice reduction, amplified by moisture increase. The tip is formed by relative humidity increase and wind weakening in relation to sea level pressure (SLP). In these regime-dependent, single-hemisphere processes, the atmospheric responses in SUSI plus RAD play a key role alongside surface albedo feedback, with water vapor feedback providing significant amplification. By contrast, coupled air-sea interaction and cloud feedback, which are often invoked to explain regional SST change, are excluded.

### **Forced circulation fingerprints on the hook-shaped warming**

The Coupled Model Intercomparison Project (CMIP) ensemble mean ([Methods](#)) presents SST warming in the shape of a hook ([Fig. 1b](#)), as highlighted by the green arrow in [Fig. 1c](#), which curves around the climatological NPSA (see [Fig. 1a](#)). At the hook tip (orange parallelogram) south of the NPSA, subtropical trade winds weaken with southwesterly anomalies, suppressing evaporative cooling and favoring relative SST warming. This wind pattern ([Fig. 1c](#)) also occurs in SUSI plus RAD run with atmosphere-only models ([Fig. 1a](#)), leading to a very high spatial correlation ( $r_s$ ) of 0.9 between the corresponding SLP responses. Thus, the circulation change driving the hook-tip warming is an atmospheric response to CO<sub>2</sub> effect plus uniform warming, in contrast to a response to local SST anomalies ([Fig. 1b](#)).

Throughout the scenarios in Fig. 1, surface wind anomalies closely follow the SLP change<sup>12</sup>, indicating the dominance of quasi-geostrophic balance (Methods). The trade wind weakening at the hook tip is part of the SUSI-induced reduction in the NPSA, with negative SLP anomalies (Extended Data Fig. 1a) opposing the climatology (Fig. 1a). The NPSA is characterized by a persistent high-pressure center near 30°N, maintained by monsoon systems, land-sea thermal contrast, and large-scale subsidence<sup>23-25</sup>. Its weakening in the SUSI experiment is driven by the mean advection of stratification change (MASC) mechanism<sup>26</sup>: tropospheric warming increases with height, so that climatological subsidence advects more warmed air downward, producing adiabatic heating to weaken the circulation system (see Methods).

In SUSI, the NPSA reduction connects to a strengthened Aleutian Low<sup>27</sup> (Extended Data Fig. 1a), but an anomalous anticyclone induced by RAD<sup>28-31</sup> (Extended Data Fig. 1b) weakens this connection (see Fig. 1a). The Aleutian Low intensification enhances mid-latitude winds along the hook string (blue trapezoid) from southwest to northeast (Fig. 1c). Although this could potentially cool the SST through evaporation, the actual warming is so rapid and intense that extremely hot summers have recently been experienced in Northeast Asia<sup>32</sup>. Hence, the string must arise from other factors, such as radiative effect and surface stability in turbulent fluxes. A heat budget analysis of the ocean mixed layer is required to identify these key factors.

In turn, the hook-shaped  $T^*$  causes adjustments to the SLP and associated winds (Fig. 1b), slightly enhancing the Aleutian Low but partly offsetting the NPSA reduction. The diminished wind weakening on the southwestern and northeastern NPSA flanks probably produces the open boundary of the hook (Fig. 1c). Hook-tip warming reduces the local SLP and further weakens trade winds south of the tip. Nevertheless, these adjustments are small in magnitude, indicating that air-sea decoupling reduces the importance of WES in forming the North Pacific SST warming pattern. To quantify this, we incorporate a heat budget analysis of the ocean mixed layer and establish our response-feedback-interaction framework.

## Regime distinction according to the heat budget analysis

To examine the hook-shaped warming, we employ a mixed-layer heat budget (see [Methods](#)), which breaks down the SST pattern into components attributable to latent and sensible (turbulent) heat fluxes, radiative effect, and oceanic transport. We then divide the turbulent heat changes by wind speed, surface stability (the deviation of SAT from SST) and relative humidity (for latent heat only). Such bulk-flux decompositions are widely used to diagnose air-sea fluxes<sup>33,34</sup>, and we convert them into SST anomalies to quantify their relative contributions<sup>1,12,13,27,35</sup>. Under 1pctCO2, the sum of the components ([Extended Data Fig. 2a](#)) is almost identical ( $r_s = 0.97$ ) to the projected  $T^*$ , validating our decomposition (since the latent-heat stability effect requires approximation for feasibility).

A clear regime distinction separates the hook string and tip. The mid-latitude string (around 25°-60°N) warms significantly, with a spatial mean of 0.35 K K<sup>-1</sup>, whereas the subtropical tip (about 5°-25°N) warms moderately, at 0.051 K K<sup>-1</sup> ([Extended Data Fig. 2a](#)). The strongest factor to the string is radiative effect ([Extended Data Fig. 2b](#);  $r_s = 0.9$  with the total SST). This may be stimulated by the sensible surface stability ([Extended Data Fig. 2c](#);  $r_s = 0.66$  over 5°-50°N), plus its latent counterpart ([Extended Data Fig. 2d](#);  $r_s = 0.55$  with the sensible). Despite the enhanced vector winds associated with the Aleutian Low intensification ([Fig. 1c](#)), the corresponding change in wind speed is too weak (not shown) to evidently cool the ocean, probably due to the non-linearity of its calculation. Instead, the speed effect in the subtropics influences the hook tip through latent heat (see [Extended Data Fig. 2f](#)), together with relative humidity ([Extended Data Fig. 2h](#)). By contrast, sensible wind speed and oceanic flux ([Extended Data Fig. 2e, g](#)) have little impact on the total SST ( $r_s = -0.2$  and 0.11, respectively).

The hook-shaped warming pattern is therefore not attributed to a single basin-scale mechanism, but to multiple processes. The string requires tracing the stability effect and radiative enhancement, while the tip needs an explanation of the latent heat terms. For these, we establish our response-feedback-interaction framework by isolating SUSI and RAD in the heat budget. This approach is typically applied to coupled runs with a closed budget, such as 1pctCO2. When the SST is prescribed, the mixed layer cannot respond

to heat fluxes, leaving the budget unbalanced. In our SST-converted analysis, the heat fluxes in SUSI and RAD are treated as potential perturbations. If present in a coupled model, they would produce initial  $T^*$  responses to trigger radiative feedbacks and/or air-sea interactions. This distinguishes the initial atmospheric (SUSI and RAD) effect from the subsequent  $T^*$ -driven effect.

#### **Hook string: surface stability effect due to warm-air advection**

For the key factors, Fig. 2 breaks the heat budget down by scenario, with the  $T^*$ -driven effect diagnosed as the coupled changes (1pctCO2) minus the atmospheric responses (SUSI plus RAD). Consisting of sensible and latent components (Extended Data Fig. 2c, d), the stability effect (Fig. 2a) originates from greenhouse gas forcing (Fig. 2b) around the Eurasian continent to the west and the high latitudes to the east ( $r_s = 0.53$ ), and then propagates into the North Pacific interior (Fig. 2c) by air-sea interaction ( $r_s = 0.71$ ). This is regulated by SAT change (Fig. 3) relative to SST: a faster warming of the lower troposphere reduces the air-sea temperature difference, increasing surface stability and suppressing turbulent heat release, which relatively warms the SST.

The excessive SAT warming (Fig. 3a) is thus traced to near-surface horizontal atmospheric advection (Extended Data Fig. 3a-c), which well explains the surface stability effect in Fig. 2a-c ( $r_s = 0.47, 0.49$  and  $0.48$ , respectively). This heat transport is further decomposed into thermodynamic and dynamic components (Methods). Under the 1pctCO2 and  $T^*$ -driven scenarios (Extended Data Fig. 3a, c), the thermodynamic effect is more significant ( $r_s = 0.81$  and  $0.87$ ) than the dynamic ( $r_s = 0.67$  and  $0.49$ , respectively). Whereas in SUSI plus RAD (Extended Data Fig. 3b), the dynamic effect dominates ( $r_s = 0.74$ ) over the thermodynamic ( $r_s = 0.62$ ; blocked by prescribed SSTs).

The thermodynamic component describes anomalous SAT advected by climatological winds (Extended Data Fig. 3d). It produces the western coastal warming in SUSI plus RAD (Extended Data Fig. 3e), with comparable contributions (see Supplementary Figs. 1 and 2). Owing to weaker evaporative damping<sup>36-39</sup>, SAT warming is robustly stronger over land than over ocean (Fig. 3b). This is advected

seaward by the mean westerlies, but blocked by the prescribed SSTs that artificially uptake all SAT heating. Once the SST prescription is released in the  $T^*$ -driven results (Extended Data Fig. 3f), the warm advection extends into the northwestern North Pacific (Fig. 3c). Yet, this is not actually a coupled air-sea interaction, but merely the forced response of SST to SAT without the artificial, limitless heat sink that exists in atmosphere-only simulations.

Conversely, the dynamic component represents climatological SAT advected by anomalous winds (Extended Data Fig. 3g). Supplementary Figs. 1 and 2 attribute the warming in the eastern high-latitude and western off-coast regions (Extended Data Fig. 3h) to SUSI and RAD through the local southeasterlies and southerlies (Extended Data Fig. 1a, b), respectively. These both transport tropical hot air towards the cold extratropics (Fig. 3e). The thermodynamic and western dynamic effects then drive an initial southeastward SLP gradient, directing southwesterlies off the Japanese coast. This brings more tropical heat to warm the ocean, thereby extending the SLP gradient and southwesterlies northeastwards (Fig. 3f). This air-sea interaction propagates the western SAT warming into the North Pacific interior (Extended Data Fig. 3i), to form the hook string and associated circulation change in Fig. 1b.

## **Radiative effect: albedo and water vapor feedbacks**

The radiative effect (Fig. 2d) is almost absent for SUSI plus RAD (Fig. 2e), but appears under  $T^*$  driving (Fig. 2f;  $r_s = 0.97$  with Fig. 2d). Fluxes in clear and cloudy skies (Supplementary Fig. 3) lead to similar ( $r_s \geq 0.94$ ) SST decomposition (Fig. 4), suggesting the dominance of radiative perturbations<sup>40</sup> for the  $T^*$  component. Surface downward shortwave effect (Fig. 4a) caused by cloud change (Fig. 4c;  $r_s = 0.78$ ) is weak. Instead, the upward shortwave and downward longwave effects dominate (Fig. 4d, g;  $r_s = 0.92$  and  $0.98$  with Fig. 2f, respectively), which are governed by clear-sky changes (Fig. 4e, h;  $r_s > 0.97$ ).

The shortwave effect is controlled by upward (Fig. 4e) rather than downward (Fig. 4b) flux, indicating reduced solar reflection at the sea surface. This occurs only in winter

(Extended Data Fig. 4a), implying a clear-sky albedo reduction (Extended Data Fig. 4b) caused by melting sea ice ( $r_s = -0.99$ ). Both SST and albedo changes collocate with high climatological albedo (Extended Data Fig. 4c;  $r_s = 0.94$  and  $-0.92$ , respectively), within the freezing line of winter SST (Extended Data Fig. 4d). Here, SST warming reduces sea-ice concentration, allowing ocean to absorb more solar radiation and warm further<sup>40</sup>. This clear-sky albedo feedback<sup>41</sup> is intense north of 45°N but absent in summer, when climatological sea ice melts completely (Extended Data Fig. 4e-h). It also disappears in SUSI plus RAD (Fig. 2e), which lacks the response of sea ice to warming.

The clear-sky downward longwave radiation instead covaries with atmospheric moistening over relatively warmed SSTs, covering a broad mid-latitude band of the hook string. The vertically averaged relative humidity change (Fig. 5a) correlates with  $T^*$  ( $r_s = 0.48$ ), indicating convective adjustment for precipitation change (Fig. 5b;  $r_s = 0.8$  with Fig. 5a). It increases specific humidity (Fig. 5c;  $r_s = 0.85$ ,  $0.74$  with Fig. 5a, b) and the downward longwave flux (Supplementary Fig. 3h;  $r_s = 0.76$ ), to warm the SST again. This constructs a convective water vapor feedback loop, incorporating a thermodynamic feedback from increased vapor-holding capacity due to relative atmospheric warming. Note that near-surface air warming may further strengthen downward longwave emission<sup>42</sup>. Converted into SST contribution (Fig. 4h), the poleward reduction in the Newtonian plus Plank damping (Methods) lowers the moisture correlation to 0.6. This renders<sup>40</sup> the flux maximum at the hook tip ineffective in driving the SST, blocking the local water vapor feedback pathway.

Our response-feedback-interaction framework thus fully addresses the hook string formation. The intensification of the Aleutian Low has little impact on the SST, as the wind speed change and its cooling effect are insignificant (see Fig. 2g and Extended Data Fig. 2e, f). This gives rise to the following processes. Under 1pctCO<sub>2</sub>, the thermodynamic stability effect involves zonal warm advection (Extended Data Fig. 5d) via climatological westerlies ( $r_s = 0.6$ ), due to an increase in land-sea thermal contrast (Fig. 3a). The dynamic effect utilizes meridional warm advection (Extended Data Fig. 5e) by anomalous southerlies ( $r_s = 0.66$ ), based on the tropical-extratropical temperature

difference in climatology (Fig. 3d). The albedo feedback is intense but confined to the northwest. These warming effects combine to form an initial pattern, which triggers strong water vapor feedback extending across the entire hook string ( $r_s = 0.88$ ). The trigger and feedback processes contribute comparably to the total SST (Extended Data Fig. 2a), with  $r_s$  of 0.85 and 0.9, respectively.

#### **Hook tip: relative humidity increase related to precipitation change**

The hook tip is mainly controlled by latent heat through wind speed and relative humidity changes (Extended Data Fig. 2f, h), which contribute to spatial-mean (5°-25°N) warming of 0.025 and 0.026 K K<sup>-1</sup>, respectively. They are mainly associated with forced responses to large-scale NPSA and its change, rather than cross-equatorial air-sea interaction. As mentioned before, anomalous southwesterlies weaken the northeasterly trade winds (Fig. 1c) and reduce evaporative cooling. In Fig. 2g, this wind speed effect largely comes from SUSI plus RAD (Fig. 2h), with a minor contribution from air-sea interaction (Fig. 2i). These are due to the SUSI-induced NPSA reduction (Fig. 1a) and the  $T^*$ -driven low pressure anomaly (Fig. 1b), respectively. The dominance of the forced response and local quasi-geostrophy rejects the concept of cross-equatorial WES feedback for the tip warming.

An increase in relative humidity<sup>43,44</sup> reduces the moisture contrast between the surface ocean and overlying atmosphere, again weakening evaporative cooling (Fig. 2j). It is primarily evident in SUSI plus RAD (Fig. 2k), particularly in SUSI (Supplementary Fig. 1g), whereas the  $T^*$ -driven contribution is negative in the tip region (Fig. 2l). Consistently, in the meridional mean for 15°-25°N, the relative humidity increase (Supplementary Fig. 4) is dominated by SUSI in the mid- and lower troposphere (Supplementary Fig. 4c). We therefore examine the hydrological cycle change in SUSI, which supplies net moisture (see Methods) to Extended Data Fig. 6a. Here, evaporation minus precipitation ( $E-P$ ) enhances (Extended Data Fig. 6b) due to suppressed rainfall (Extended Data Fig. 6c). This is governed by the dry-get-drier paradigm associated with atmospheric moistening in greenhouse warming<sup>3,4</sup>, since the tip lies beneath prevalent large-scale NPSA subsidence (see Extended Data Fig. 6d). Consequently, the relative

humidity increase correlates well with the climatological 500-hPa pressure velocity and precipitation change ( $r_s = 0.65$  and  $-0.45$ , respectively).

### **Spatial applicability and uncertainty attribution**

According to our response-feedback-interaction framework, the two regimes of North Pacific SST warming are shaped by distinct combinations of albedo (Fig. 4e) plus water vapor (Fig. 4h) feedback, surface stability (Fig. 2a), wind speed (Fig. 2g), and relative humidity (Fig. 2j) changes. The last two form the hook tip, while the first three overturn the weak wind speed change associated with the Aleutian Low, thereby warming the string. The former two embrace the radiative effects, and the latter three the atmospheric responses (SUSI plus RAD). Fig. 6 illustrates their spatial applicability in ensemble mean: the albedo, stability and water vapor effects shift from the northwestern boundary towards the interior of the North Pacific, while wind speed and relative humidity control the eastern subtropics together, with their influences extending southwestward and northwestward, respectively.

Extended Data Fig. 7 attributes the intermodel uncertainty in the 1pctCO<sub>2</sub> SST pattern (Extended Data Fig. 2) to the key factors. Those responsible for the hook tip and string comparably account for 46% and 48% respectively and 94% in sum, as calculated<sup>12</sup> (Methods) from intermodel correlations of 0.7, 0.72, and 0.97 (Extended Data Fig. 7a). The tip variance (Extended Data Fig. 7b) is mainly due to wind speed (80%) rather than relative humidity (20%). Surface stability and albedo effects explain 25% and 22% of the string variance (Extended Data Fig. 7c), comparably amplified by water vapor feedback (53%). Overall, the atmospheric responses and radiative feedbacks contribute similarly to the total, at 54% and 40%, respectively (Extended Data Fig. 7d).

In summary, North Pacific SST pattern formation is attributed to single-hemisphere, regime-dependent mechanisms, rather than to a cross-equatorial process. While the atmospheric responses provide an underlying  $T^*$  structure, albedo plus water vapor feedback exerts significant enhancement in the mid-latitudes. The trigger and

feedback loop depends on the surface stability response and clear-sky radiative effect,  
instead of air-sea interaction or cloud change commonly considered to drive SST  
anomalies. Consequently, the hook string warming is markedly strong and robust, yet  
the tip can be cooled by the internal variability through wind change<sup>45</sup>. Clarifying this  
surprising and complicated process is a key novelty here.

## Discussion

This study presents a new process-based framework for SST pattern formation in the North Pacific, which distinguishes between the atmospheric responses, radiative feedbacks, and air-sea interactions. It is established by combining scenario decomposition with a mixed-layer heat budget and incorporating physical diagnostics of quasi-geostrophy, surface air advection, albedo, and water vapor. Although the northern hook and southern SHIW<sup>12</sup> presents apparently related regional warming patterns, they arise from distinct climate processes. For example, shortwave albedo feedback is unlikely in 45°-60°S, given the ~2 K warmer climatological SST than in 45°-60°N for 100°E-70°W. In addition, the atmospheric responses contribute to trigger water vapor feedback, causing significant longwave amplification under a clear sky. In the South Pacific, these merely provide damping<sup>12</sup>, being overturned by strong air-sea interaction and cloud feedback. The surprisingly marked contrast may stem from an asymmetry in tropical expansion<sup>46,47</sup> (the trigger of SHIW), which is generally stronger in the Southern Hemisphere.

In the Northern Hemisphere, tropical expansion is regional and seasonal<sup>48,49</sup>. In addition to the zonal asymmetry of the surrounding continents, the land-sea thermal contrast involves competing tendencies. In particular, the direct radiative effect promotes poleward motion<sup>38,39</sup>, while the SST warming (e.g., the hook string) favors equatorward adjustment<sup>31</sup>. Consequently, the Pacific tropics experience an equatorward contraction in summer, leaving the expansion to the Atlantic and winter, rather than composing a zonal or annual-mean poleward shift<sup>24,28,29,50</sup>. These distinct initial conditions therefore steer the SST pattern formation through contrasting pathways between the hemispheres. Ultimately, constraining warming uncertainty in the North Pacific depends on simulation skills for SAT, NPSA, humidity and sea ice. Furthermore, our framework could be utilized in other ocean basins, to test the applicability of the northern response-feedback mechanism against the southern air-sea interaction process.

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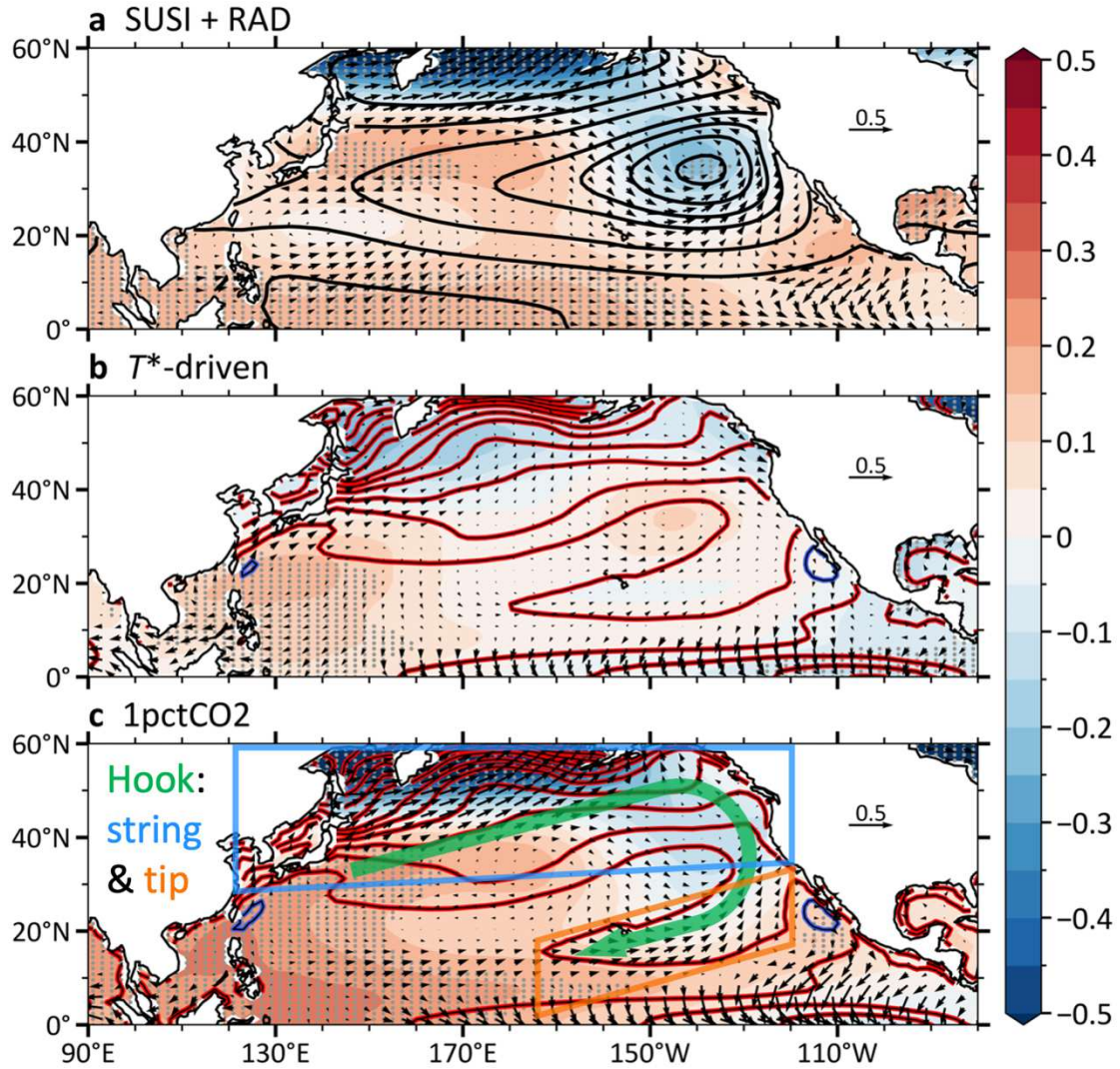


**Fig. 1 | Ensemble-mean changes in SLP (shading,  $\text{hPa K}^{-1}$ ) and surface winds (vectors,  $\text{m s}^{-1} \text{K}^{-1}$ ) under the (a) SUSI plus RAD, (b)  $T^*$ -driven, and (c) 1pctCO2 scenarios. The background contours show the present-day SLP [contour interval (CI): 1.5 hPa above 1018 hPa and 3 hPa below 1015 hPa] in (a), and the SST warming pattern (CI: 0.1  $\text{K K}^{-1}$  below 0.55  $\text{K K}^{-1}$  and 0.2  $\text{K K}^{-1}$  above 0.75  $\text{K K}^{-1}$ ; red positive, blue negative; zero omitted) in (b, c). Stippling indicates that the ensemble mean exceeds the intermodel spread. The SLP and surface winds follow the quasi-geostrophic balance in all scenarios, and the warming pattern exhibits a hook-like shape (green curved arrow in c) around the NPSA, with its tip (orange parallelogram) anchoring weakened winds south of the high, and its string (blue trapezoid) extending from southwest to northeast.**

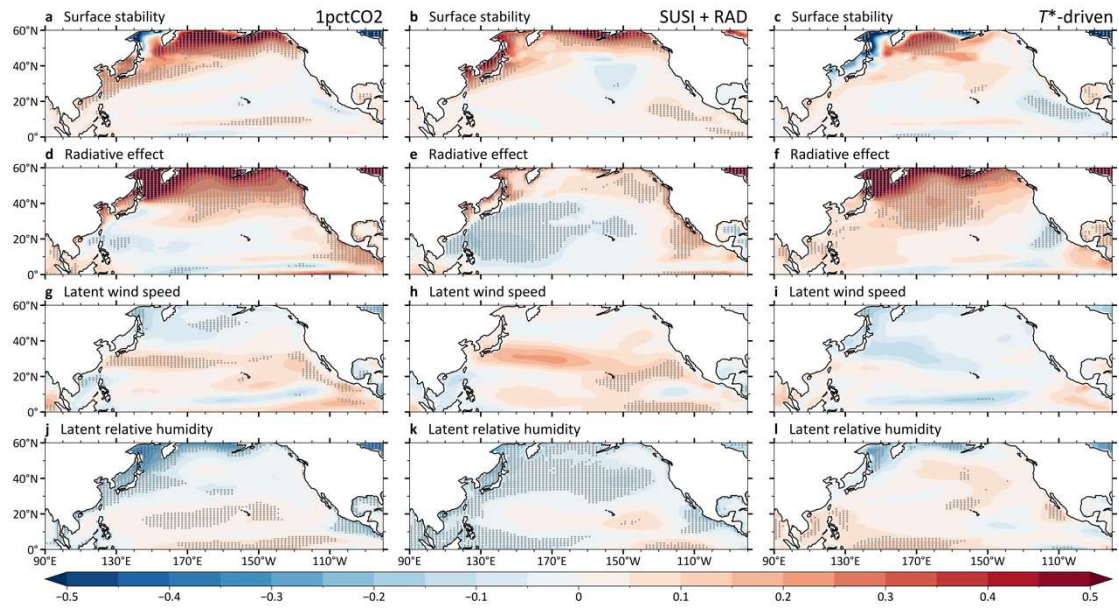


**Fig. 2 | Heat budget decomposition (see [Methods](#)) of the SST warming pattern ( $K$   $K^{-1}$ ) for diagnosing the key factors across all scenarios.** Each row shows one factor: (a-c) surface stability, (d-f) radiative effect, (g-i) latent wind speed, and (j-l) latent relative humidity – for the coupled 1pctCO2 projection (left), forced contribution from SUSI plus RAD (middle), and  $T^*$ -driven effect (right). Stippling denotes that the ensemble-mean SST component exceeds its intermodel spread, and a positive value represents ocean warming. Surface stability effect appears in both SUSI plus RAD and the  $T^*$ -driven projection with distinct spatial patterns, the radiative effect are mainly driven by  $T^*$ , and latent wind speed and relative humidity are primarily forced responses.

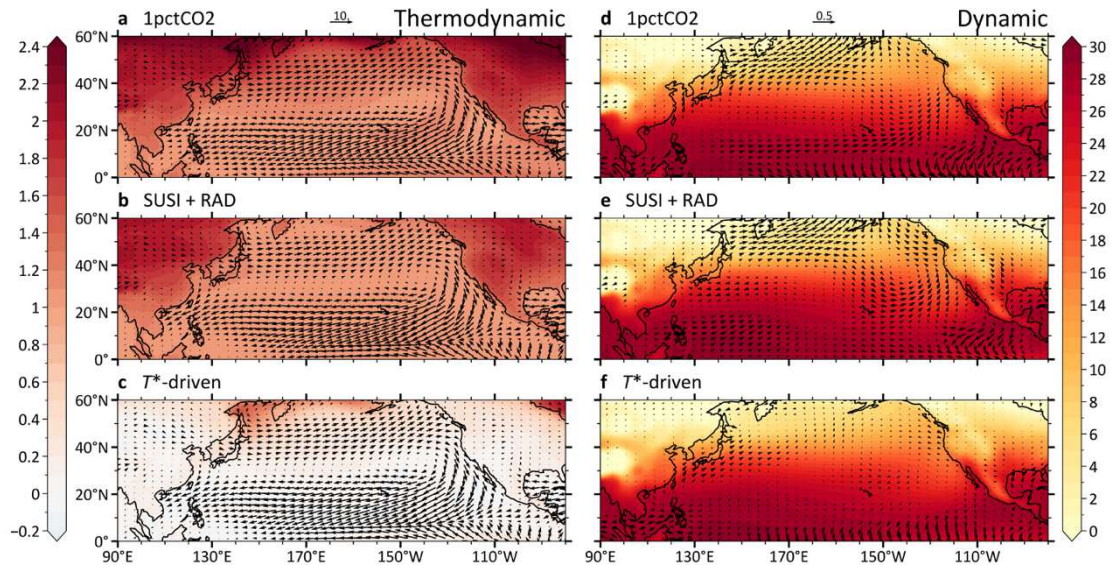


**Fig. 3 | Decomposition elaborating the SAT advection over the hook string for the**  
**(a, d) 1pctCO2 projection, (b, e) SUSI plus RAD, and (c, f)  $T^*$ -driven effect,**  
**including the thermodynamic (a-c) and dynamic (d-f) components.** The former is  
defined as SAT changes (shading,  $\text{K K}^{-1}$ ) advected by climatological winds (vectors,  $\text{m s}^{-1}$ ), and the latter as climatological SAT (shading,  $^{\circ}\text{C}$ ) advected by wind changes  
(vectors,  $\text{m s}^{-1} \text{ K}^{-1}$ ). The dynamic component reflects meridional advection by  
anomalous southerlies transporting the tropical hot atmosphere northward, whereas the  
thermodynamic effect represents zonal advection by climatological westerlies  
extending the more warmed land air seaward.

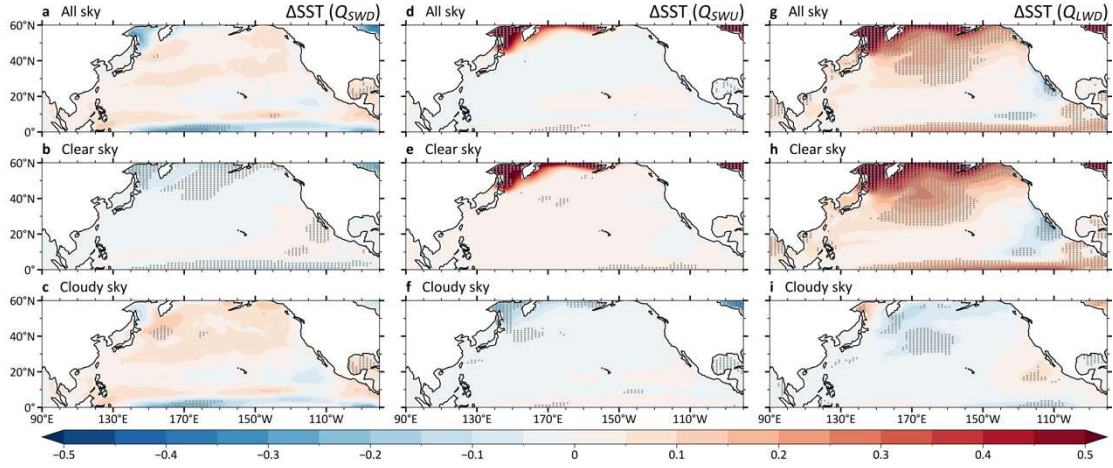


**Fig. 4 | SST decomposition for the radiative effect ( $\text{K K}^{-1}$ ) in the  $T^*$ -driven projection, for (a-c) downward shortwave  $\Delta\text{SST} (Q_{SWD})$ , (d-f) upward shortwave  $\Delta\text{SST} (Q_{SWU})$ , and (g-i) downward longwave  $\Delta\text{SST} (Q_{LWD})$  radiation. The rows separate all (upper), clear (middle), and cloudy (lower) skies. A positive value represents ocean warming, and stippling indicates that the ensemble mean exceeds the intermodel spread. All panels are ensemble means over the 19 models providing clear-sky output, including the all-sky ones, so that the three sky conditions remain additive. The radiative effect is dominated by two clear-sky components:  $Q_{LWD}$  warms the hook string across the mid-latitudes, whereas  $Q_{SWU}$  contributes mainly north of  $45^\circ\text{N}$ .**

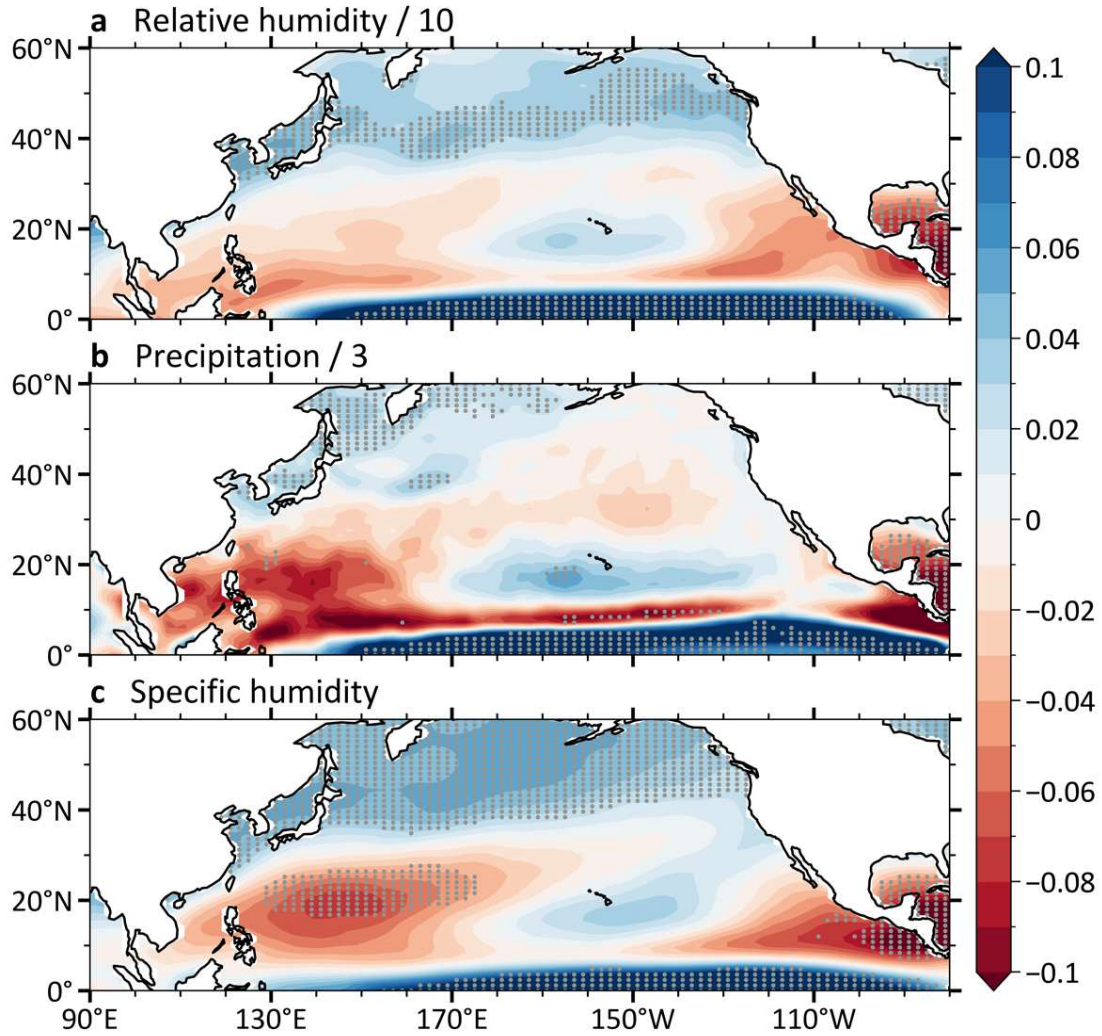


**Fig. 5 | Moisture and rainfall changes in the  $T^*$ -driven effect (shading), including the mass-weighted vertical-averaged changes (1000-100 hPa) in (a) relative humidity ( $\% \text{ K}^{-1}$ ) and (c) specific humidity ( $\text{g kg}^{-1} \text{ K}^{-1}$ ), as well as (b) precipitation change ( $\text{mm day}^{-1} \text{ K}^{-1}$ ). Relative humidity change are divided by 10 and precipitation change by 3 to share the color scale. Stippling indicates that the ensemble mean exceeds the intermodel spread. An increase in moisture and precipitation follows the SST pattern, indicating that convective water vapor feedback moistens the atmospheric column and provides clear-sky downward longwave warming.**

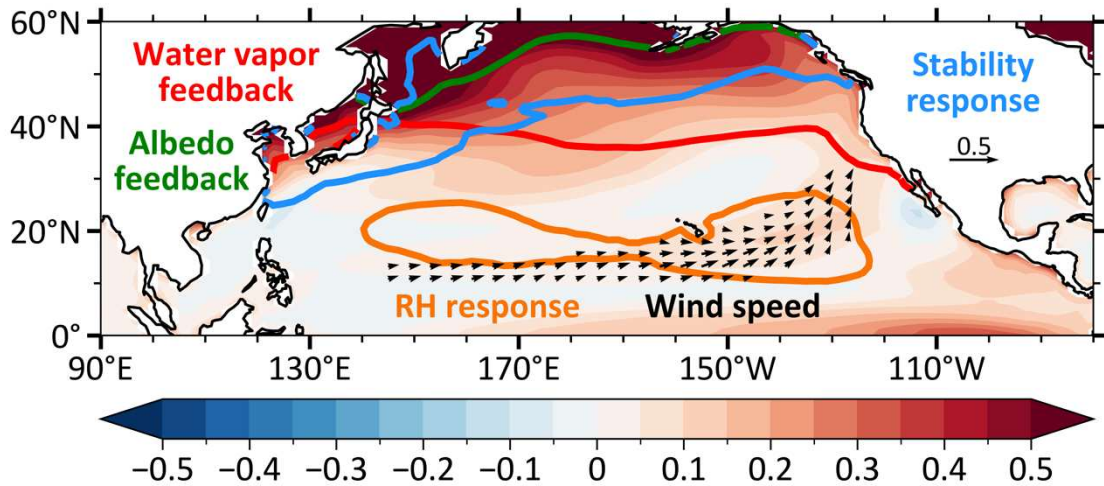


**Fig. 6 | Spatial synthesis of the dominant processes shaping SST warming pattern (shading,  $\text{K K}^{-1}$ ) in the North Pacific.** The key heat-budget factors are marked by vectors for surface wind speed change beyond  $0.1 \text{ m s}^{-1} \text{ K}^{-1}$ , and by contours for surface stability response (blue), albedo (green) and water vapor feedbacks (red) at  $0.15 \text{ K K}^{-1}$ , and relative humidity (RH) response (orange) at  $0.2 \% \text{ K}^{-1}$ . These thresholds marking the governing regions are reproduced from Figs. 1c, 2a, 4e, and 4h, and Extended Data Fig. 6a, respectively. The albedo, stability and water vapor effects extends from the northwestern side towards the entire string, whereas relative humidity and wind speed dominate the tip together.

## Methods

**Multi-model CMIP data and scenario decomposition.** We use monthly outputs from 20 CMIP models, including 11 CMIP5 models and 9 CMIP6 models<sup>51,52</sup>. The experiments include the 1% CO<sub>2</sub> increase per year experiment (1pctCO<sub>2</sub>), the Atmospheric Model Intercomparison Project (AMIP), the uniform 4 K SST warming experiments (AMIP4K in CMIP5 and AMIP-p4K in CMIP6), and the fixed-SST quadrupled CO<sub>2</sub> experiment (AMIP-4×CO<sub>2</sub>). The analysed variables include SST, SLP, SAT, precipitation, specific humidity, relative humidity, surface wind speed and vectors, pressure velocity, geopotential height, surface sensible, latent heat, and radiative fluxes. When a target variable is unavailable, we use physically consistent substitutes or derived quantities where possible, and exclude a model for a given diagnostic only when neither the variable nor a suitable alternative exists.

The CMIP5 models include BCC-CSM1.1, CCSM4, CNRM-CM5, CanESM2, HadGEM2-ES, IPSL-CM5A-LR, IPSL-CM5B-LR, MIROC5, MPI-ESM-LR, MPI-ESM-MR, and MRI-CGCM3. The CMIP6 models include BCC-CSM2-MR, CNRM-CM6-1, CanESM5, E3SM-1.0, GFDL-CM4, GISS-E2-1-G, HadGEM3-GC31-LL, IPSL-CM6A-LR, and MIROC6.

All fields are interpolated to a common 1.2° latitude × 2° longitude grid before analysis. The coupled change is defined from 1pctCO<sub>2</sub> as the difference between the 20-year means of model years 121-140 and 1-20. In our scenario decomposition<sup>12</sup>, the uniform SST warming (SUSI; AMIP4K or AMIP-p4K minus AMIP) and the direct CO<sub>2</sub> effect (RAD; AMIP-4×CO<sub>2</sub> minus AMIP) are summed to give the forced responses, and the residual between 1pctCO<sub>2</sub> and SUSI plus RAD is taken as the  $T^*$ -driven component including radiative feedbacks and air-sea interactions. All changes are normalized by the extended tropical-mean SST warming over 40°S-40°N before calculating ensemble statistics; for RAD, the normalization factor<sup>12</sup> is taken from the SST warming between piControl and abrupt-4×CO<sub>2</sub>. Relative SST warming patterns are obtained by subtracting the extended tropical mean from the normalized SST

warming, isolating the pattern from the spatially uniform warming component.

**Intermodel statistical analysis.** Regular statistics use spatio-temporal fields as inputs, but here we replace the time axis with model series to perform intermodel statistical analysis. In particular, intermodel correlations and regressions are performed between multiple variables such as SST warming patterns and their heat-budget components, and variances explained by different factors are estimated as the novel  $R^2$  for the total effect below. Student's  $t$ -tests are performed to assess the confidence of the correlations with the following formula,

$$t_r = \frac{\sqrt{n-3}}{2} \ln \left( \frac{1+r}{1-r} \right), \quad (1)$$

where  $t_r$  is the test statistic derived from the correlation ( $r$  here or  $r_s$ ) and sample size  $n$ . All  $r$  across 20 models are above 90% confidence level, and  $r_s$  pass 99.99% confidence tests.

We adopt the novel  $R^2$  method<sup>12</sup> to solve the overfitting problem of linearly interdependent factors in uncertainty attribution. For example, the sum of the variances (direct  $R^2$ ) exceeds 1 for linearly dependent factors. To correct this, we need to remove the part that is repeatedly counted in different factors due to linear dependency.  $Z = A + B$  is the simplest case with 2 factors. The results of the novel  $R^2$  are denoted as  $R_A^2$  and  $R_B^2$ , and the direct  $R^2$  is equal to the squared correlations  $R^2(Z, A)$  and  $R^2(Z, B)$  for  $A$  and  $B$ , respectively. The repetitive part between  $A$  and  $B$  is excessively counted by twice ( $2 \times 1$ ) in  $R^2(Z, A) + R^2(Z, B)$ , so that it is expressed as  $R_{AB}^2 = \frac{R^2(Z,A)+R^2(Z,B)-1}{2 \times 1}$ .

With the sum guaranteed to be 1, the novel  $R^2$  can be obtained as,

$$\begin{cases} R_A^2 = R^2(Z, A) - R_{AB}^2 = \frac{1+R^2(Z,A)-R^2(Z,B)}{2} \\ R_B^2 = R^2(Z, B) - R_{AB}^2 = \frac{1-R^2(Z,A)+R^2(Z,B)}{2} \end{cases} \quad (2)$$

**Atmospheric dynamical background.** The WES feedback<sup>17,18</sup> describes an ocean–atmosphere interaction that generates a cross-equatorial antisymmetry: a northern warm SST perturbation causes cross-equatorial southerly winds; the Coriolis force then deflects the winds to slow down the trades north of the equator and speed up the trades

south; the evaporation of the northern tropics is weakened, thereby warming the local SST, and vice versa. This positive feedback can amplify the initial disturbance and cause the intertropical convergence zone to shift northwards from the equator. It has also been proposed to form a south-to-north gradient of SST warming in the subtropical Pacific<sup>2</sup>.

Since its assumptions deviate from the quasi-geostrophic balance, WES is usually defined within a latitudinal band called the equatorial Rossby radius of deformation. This radius defines the length scale on which the Coriolis force is comparable to the acceleration and pressure gradient force<sup>53</sup>. Equatorial wave dynamics determine its magnitude as  $R = \sqrt{C_0/\beta} \sim 10^\circ\text{-}15^\circ$  for the atmosphere, where  $C_0 \sim 50$  m s<sup>-1</sup> is the atmospheric wave speed, and  $\beta = df/dy = 2\Omega \cos \theta/a \sim 2.29 \times 10^{-11}$  m<sup>-1</sup> s<sup>-1</sup> at the equator.  $f = 2\Omega \sin \theta$  is the Coriolis parameter, where  $\Omega$  is the Earth's angular speed of rotation, and  $\theta$  is latitude;  $y = a \theta$  is the latitudinal distance, where  $a$  is the mean radius of the Earth. The circulation systems examined here — the NPSA and the Aleutian Low — lie well poleward of this range, so that the SLP and wind changes analysed in this study follow the quasi-geostrophic balance rather than equatorial wave dynamics.

Outside the equatorial Rossby radius of deformation, the large-scale atmospheric circulation is governed by the quasi-geostrophic balance. The geostrophic wind  $\mathbf{V}_g$  is expressed as  $\mathbf{V}_g = \mathbf{k} \times \nabla p/(\rho f)$ , where  $\mathbf{k}$  is the unit  $z$  vector,  $\nabla$  is the horizontal gradient,  $p$  is pressure, and  $\rho$  is air density. Here,  $-\nabla p/\rho$  is defined as the pressure gradient force in balance with the Coriolis force.  $\mathbf{V}_g$  is perpendicular to the pressure gradient force and therefore parallel to the isobars. In fact, the actual wind is determined by a 3-force balance with the addition of friction, so that it is close to the geostrophic wind, but with a relatively weak strength and a direction tilted towards low pressure.

With respect to climate change, the global-mean water vapor and vertical gradient of air temperature increase at  $\sim 7\%$  K<sup>-1</sup> of surface warming, but the fractional

changes in global mean precipitation and net longwave radiative cooling are only 1-2% K<sup>-1</sup> based on model projections. This contrast indicates a weakening of the circulation at a rate of ~5% K<sup>-1</sup> in both convective and subsidence regions<sup>3,4,54,55</sup>. The MASC is a dynamical mechanism proposed to explain the weakening of the tropospheric circulation throughout the tropics<sup>26</sup>. Following the moist adiabats (the function of atmospheric temperature decreasing with altitude), the tropics-wide warming enhances with height and increases the dry stability of the atmosphere. This vertical gradient leads to relative cooling of the air column in ascending regions because of the upward advection of cooler lower-level air, and relative warming in subsidence regions because of the downward advection of warmer upper-level air. This adiabatic heating reduces the air temperature and pressure gradients between the ascent and descent regions, and it opposes and slows down the climatological circulation.

**Mixed-layer heat budget.** To establish the response-feedback-interaction framework and quantify the contributions of different processes to the SST warming patterns, we adopt an equilibrium mixed-layer heat budget that can decompose how surface flux changes affect SST. Following previous studies<sup>1,12,13,27,35</sup>, the mixed-layer heat balance over the ocean under climate change can be written as follows,

$$\Delta Q_E + \Delta Q_H + \Delta Q_{LWD} + \Delta Q_{LWU} + \Delta Q_{SW} + \Delta D_O = 0, \quad (3)$$

where  $\Delta$  denotes the change under global warming.  $\Delta Q_{LWD}$  and  $\Delta Q_{LWU}$  are the changes in surface downward and upward longwave radiation, respectively;  $\Delta Q_{SW}$  is the change in surface net shortwave radiation;  $\Delta Q_E$  and  $\Delta Q_H$  are the changes in latent and sensible heat fluxes, respectively; and  $\Delta D_O$  represents the change in ocean heat transport, including three-dimensional ocean advection, vertical mixing, and other dynamical ocean processes. All fluxes are defined as positive downward, corresponding to ocean warming.

The latent (approximately) and sensible heat fluxes can be written as follows<sup>1,13,35</sup>,

$$Q_E \approx -\rho_a L C_E W q_s(\text{SST})(1 - RHe^{\alpha S}), \quad (4)$$

$$Q_H = \rho_a C_p C_H W S, \quad (5)$$

621 where  $\rho_a$  is the near-surface air density,  $L$  is the latent heat of vaporization,  $C_p$  is  
 622 the specific heat capacity at constant pressure,  $C_E$  and  $C_H$  are the exchange  
 623 coefficients for latent and sensible heat fluxes, respectively,  $W$  is the 10-m wind speed,  
 624  $q_s(\text{SST})$  is the saturation specific humidity at the SST,  $RH$  is the near-surface relative  
 625 humidity, and the surface stability is defined as  $S = \text{SAT} - \text{SST}$ . The coefficient  $\alpha =$   
 626  $L/(R_v \text{SST}^2) \approx 0.06 \text{ K}^{-1}$ , where  $R_v$  is the gas constant for water vapor.

627 The latent heat flux change is further decomposed into four terms:

$$628 \quad \Delta Q_E = \frac{\partial Q_E}{\partial \text{SST}} \Delta \text{SST} + \frac{\partial Q_E}{\partial W} \Delta W + \frac{\partial Q_E}{\partial RH} \Delta RH + \frac{\partial Q_E}{\partial S} \Delta S, \quad (6)$$

629 the terms on the right-hand side represent Newtonian cooling, wind-induced cooling,  
 630 and heating associated with changes in relative humidity and surface stability,  
 631 respectively. They can be written as follows,

$$632 \quad \Delta Q_{EO} = \frac{\partial Q_E}{\partial \text{SST}} \Delta \text{SST} = \alpha Q_E \Delta \text{SST}, \quad (7)$$

$$633 \quad \Delta Q_{EW} = \frac{\partial Q_E}{\partial W} \Delta W = \frac{Q_E}{W} \Delta W, \quad (8)$$

$$634 \quad \Delta Q_{ERH} = \frac{\partial Q_E}{\partial RH} \Delta RH = \frac{-Q_E}{e^{-\alpha S} - RH} \Delta RH, \quad (9)$$

$$635 \quad \Delta Q_{ES} = \frac{\partial Q_E}{\partial S} \Delta S \approx \frac{-\alpha Q_E RH}{e^{-\alpha S} - RH} \Delta S. \quad (10)$$

636 Note that the surface stability effect requires the approximation in  $\alpha$  for feasibility.

637 Similarly, the sensible heat flux change is decomposed into wind speed and surface  
 638 stability terms:

$$639 \quad \Delta Q_H = \frac{\partial Q_H}{\partial W} \Delta W + \frac{\partial Q_H}{\partial S} \Delta S, \quad (11)$$

640 with the two terms as follows,

$$641 \quad \Delta Q_{HW} = \frac{\partial Q_H}{\partial W} \Delta W = \frac{Q_H}{W} \Delta W, \quad (12)$$

$$642 \quad \Delta Q_{HS} = \frac{\partial Q_H}{\partial S} \Delta S = \frac{Q_H}{S} \Delta S. \quad (13)$$

643 The change in upward longwave radiation is approximated by the linearized  
 644 Stefan-Boltzmann relation (Plank feedback<sup>40</sup>),

$$\Delta Q_{LWU} = -4\sigma SST^3 \Delta SST, \quad (14)$$

where  $\sigma$  is the Stefan-Boltzmann constant. Because ocean heat transport changes exist only in the coupled 1pctCO2 experiment and are absent in the atmosphere-only SUSI and RAD experiments, we set

$$\Delta D_o = \begin{cases} 0, & \text{SUSI and RAD} \\ -(\Delta Q_E + \Delta Q_H + \Delta Q_{LWD} + \Delta Q_{LWU} + \Delta Q_{SW}), & \text{1pctCO2} \end{cases} \quad (15)$$

Thus, the mixed-layer heat budget can be approximately rewritten as follows,

$$\begin{aligned} & (\alpha Q_E - 4\sigma SST^3) \Delta SST + \frac{Q_E}{W} \Delta W + \frac{-Q_E}{e^{-\alpha S} - RH} \Delta RH + \frac{-\alpha Q_E RH}{e^{-\alpha S} - RH} \Delta S \\ & + \frac{Q_H}{W} \Delta W + \frac{Q_H}{S} \Delta S + (\Delta Q_{LWD} + \Delta Q_{SW}) + \Delta D_o = 0, \\ \Rightarrow \Delta SST = & \frac{\frac{Q_E}{W} \Delta W + \frac{-Q_E}{e^{-\alpha S} - RH} \Delta RH + \frac{-\alpha Q_E RH}{e^{-\alpha S} - RH} \Delta S + \frac{Q_H}{W} \Delta W + \frac{Q_H}{S} \Delta S + (\Delta Q_{LWD} + \Delta Q_{SW}) + \Delta D_o}{4\sigma SST^3 - \alpha Q_E}. \end{aligned} \quad (16)$$

This provides the response-feedback-interaction framework for decomposing the SST warming pattern. The terms on the right-hand side are the contributions from latent wind speed, latent relative humidity, latent surface stability, sensible wind speed, sensible surface stability, radiative effect, and oceanic heat transport, respectively. The denominator represents the latent Newtonian plus longwave Plank cooling. This decomposition<sup>12</sup> is applied to SUSI, RAD, and 1pctCO2, where the atmospheric responses are defined by SUSI plus RAD. The difference between 1pctCO2 and this provides radiative feedbacks (net shortwave and downward longwave terms) and air-sea interactions (turbulent flux and oceanic transport terms).

For the forced responses, both the total SST and its decomposed components are based on non-equilibrium surface fluxes with  $\Delta D_o = 0$ . They therefore represent the potential effects associated with uniform SST warming or fixed-SST radiative adjustment, which differ from these actual model projections. In the fully coupled ocean-atmosphere system, these can subsequently stimulate radiative feedbacks and/or trigger air-sea interactions. For 1pctCO2, the total reconstructed from the heat budget is almost identical ( $r_s = 0.97$ ) to the model-projected SST change. This validates the consistency of the approximate framework.

**Horizontal SAT advection.** To examine how atmospheric advection contributes to the surface stability term in the mixed-layer heat budget, we decompose the horizontal temperature advection in the lower atmosphere into thermodynamic and dynamic components. Based on the atmospheric thermodynamic equation<sup>53</sup>, the horizontal temperature advection term is written as,

$$-\mathbf{V}_h \cdot \nabla_h \text{SAT} = -u \frac{\partial \text{SAT}}{\partial x} - v \frac{\partial \text{SAT}}{\partial y}, \quad (17)$$

where  $\mathbf{V}_h$  is the horizontal wind vector,  $\nabla_h \text{SAT}$  is the horizontal air-temperature gradient,  $u$ ,  $v$  are the zonal and meridional wind components. With this sign convention<sup>53</sup>, cold advection, in which air flows from colder to warmer regions, produces local cooling and gives  $-\mathbf{V}_h \cdot \nabla_h \text{SAT} < 0$ , whereas warm advection, in which air flows from warmer to colder regions, produces local warming and gives  $-\mathbf{V}_h \cdot \nabla_h \text{SAT} > 0$ .

The change in horizontal temperature advection is decomposed as follows,

$$\begin{aligned} \Delta(-\mathbf{V}_h \cdot \nabla_h \text{SAT}) = & -(\Delta u) \frac{\partial \text{SAT}}{\partial x} - (\Delta v) \frac{\partial \text{SAT}}{\partial y} - u \frac{\partial(\Delta \text{SAT})}{\partial x} - v \frac{\partial(\Delta \text{SAT})}{\partial y} - \\ & (\Delta u) \frac{\partial(\Delta \text{SAT})}{\partial x} - (\Delta v) \frac{\partial(\Delta \text{SAT})}{\partial y}. \end{aligned} \quad (18)$$

The first two terms represent the dynamic component, in which circulation changes act on the climatological SAT gradient. The middle two terms represent the thermodynamic component, in which climatological winds advect the SAT change. The last two terms represent the nonlinear component, which is generally small compared with the dynamic and thermodynamic components and is not analysed further. Horizontal temperature advection is a key driver of lower-tropospheric temperature change, especially in the extratropics, and links circulation changes directly to the surface stability contribution to the SST warming pattern.

**Specific diagnostics of radiative effect and clear-sky albedo.** Hereby, we further investigate the specific surface fluxes to identify the pathway of the radiative effect. In particular, the net shortwave radiation ( $\Delta Q_{SW}$ ) is separated into its downward ( $\Delta Q_{SWD}$ ) and upward ( $\Delta Q_{SWU}$ ) fluxes, and analyzed alongside the downward longwave radiation

( $\Delta Q_{LWD}$ ). Clear-sky fluxes are available in the CMIP output together with the all-sky ones, so we calculate their differences as the cloudy-sky fluxes. Since the clear-sky fluxes are incomplete for GFDL-CM4, it is excluded from this analysis that uses the other 19 models.

Each component ( $\Delta Q_{XXX}$  in [Supplementary Fig. 3](#)) is then converted into its SST contribution, by following the radiative effect in [Eq. \(16\)](#) and denoted as  $\Delta SST(Q_{XXX})$  in [Fig. 4](#). The latent Newtonian plus longwave Plank cooling ( $4\sigma SST^3 - \alpha Q_E$ ) acts as a damping factor with meridional variations. Since the background SST and  $Q_E$  are larger in the tropics than in the mid-latitudes, the damping reduces poleward. Thus, a uniform ocean warming can generate increases in upward longwave radiation<sup>40</sup> and latent heat release, which reduce with latitude. This means that the SST change due to any flux would be amplified poleward by the damping effect.

The clear-sky albedo is defined as the ratio of the upward to the downward clear-sky shortwave radiation at the surface,

$$\text{Albedo (clear sky)} = \frac{Q_{SWU}(\text{clear sky})}{Q_{SWD}(\text{clear sky})}. \quad (19)$$

This is evaluated for both climatology and change. Because the sea-ice contribution is strongly seasonal, the diagnostics of albedo feedback are pursued for boreal winter and summer ([Extended Data Fig. 4](#)). Winter is taken as January-June for the fluxes and albedo, and as December-May for the SST climatology. Such a one-month lag accounts for the delayed sea-ice response to the seasonality of SST. Summer covers the rest of the months in each case.

**Relative humidity diagnostics.** To diagnose the origin of relative-humidity changes over the North Pacific, we examine the competing changes in moisture- and temperature-controlled saturation humidity<sup>3,43,44</sup>. Relative humidity is defined as  $RH = \frac{q}{q_s} \times 100\%$ , where  $q$  is specific humidity and  $q_s$  is saturation specific humidity. Changes in  $RH$  therefore reflect both atmospheric moisture content and the temperature-controlled saturation humidity: a positive response may arise from

increased moisture supply, reduced  $q_s$ , or both, whereas a negative response may reflect either moisture depletion or a faster warming-driven rise in  $q_s$ . This moisture-temperature competition is central to interpreting  $RH$  changes under global warming, since water vapor generally rises with temperature.

The moisture supply for this balance is set by the atmospheric water vapor budget. Integrating the conservation equation for specific humidity over the atmospheric column gives<sup>3,56</sup>,

$$\frac{\partial q}{\partial t} = (E - P) - \nabla \cdot q \mathbf{V}_h - \frac{\partial(\omega q)}{\partial p}, \quad (20)$$

where  $q$  is specific humidity,  $\omega$  is pressure velocity,  $E$  is surface evaporation, and  $P$  is precipitation. Evaporation minus precipitation ( $E - P$ ) is therefore the net moisture supplied by the surface to the atmosphere, while other terms represent its redistribution by the large-scale circulation. The surface acts as a moisture supply where evaporation exceeds precipitation ( $E - P > 0$ ) and as a sink where precipitation dominates ( $E - P < 0$ ).

Because this  $RH$  response reflects the moisture-temperature balance rather than a simple wind-evaporation effect, we further assess how the large-scale circulation sets the moisture supply, examining three hydrological diagnostics for the hook tip: changes in evaporation minus precipitation ( $E - P$ ), precipitation, and the climatological 500-hPa pressure velocity ( $\omega_{500}$ ) as a measure of large-scale vertical motion, with positive  $\omega_{500}$  indicating subsidence. Their spatial correspondence with the surface  $RH$  response is quantified by  $r_s$  (Extended Data Fig. 6).

## Data availability

CMIP output at phases 5 and 6 is archived at <https://esgf-node.llnl.gov> and <https://data.ceda.ac.uk/badc>.

## Code availability

The Python codes used to process and analyze data can be obtained by contacting the corresponding author.

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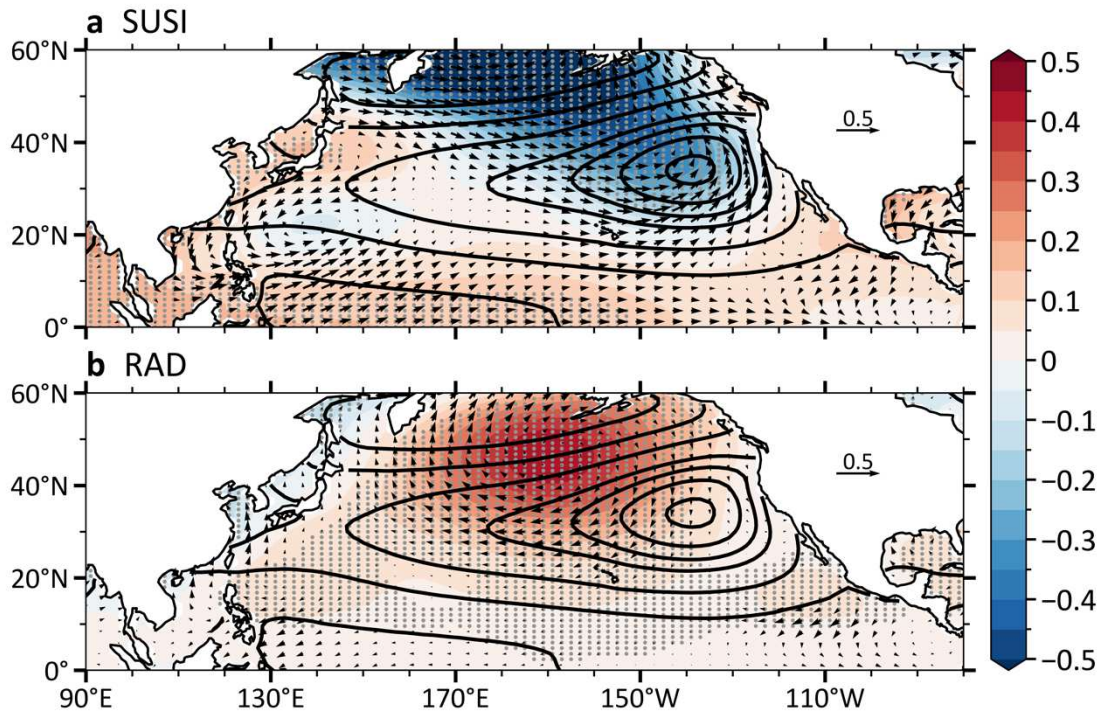
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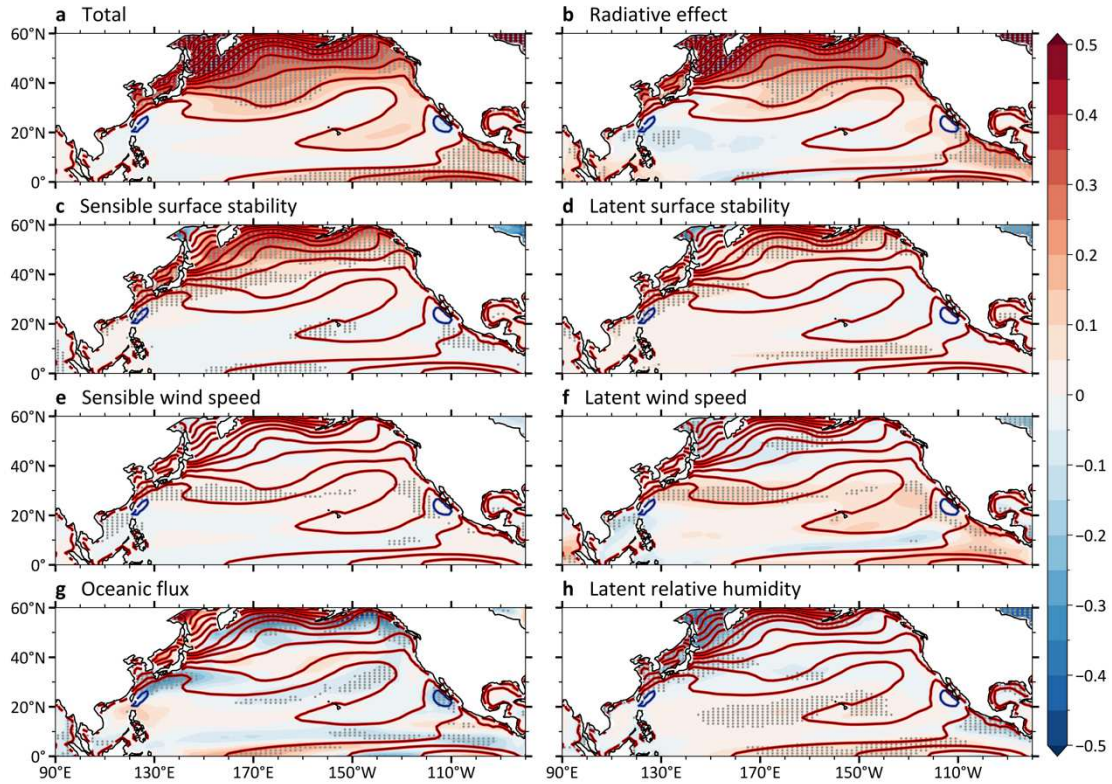
(2025YFF0516900). We thank the World Climate Research Programme, which coordinated and promoted CMIP datasets through its Working Group on Coupled Modelling. We thank the climate modelling groups for producing and making available model output, the Earth System Grid Federation (ESGF) for archiving data and providing access, and multiple agencies funding CMIP and ESGF. Helpful comments from S.-P. Xie, L. Xu, J. Ma, S. Long, C. He and L. Yang are gratefully appreciated. Python was used for analyses and graphics.

**Author contributions** X.G. and Q.S. are the co-first authors, who performed the computations and wrote the original draft. J.M. designed the study and finalized the manuscript. M.C. suggested the clear-sky diagnostics. R.C., Y.D., J.Y., C.Z., Y.L., X.Z. and E.L. revised the draft manuscript.

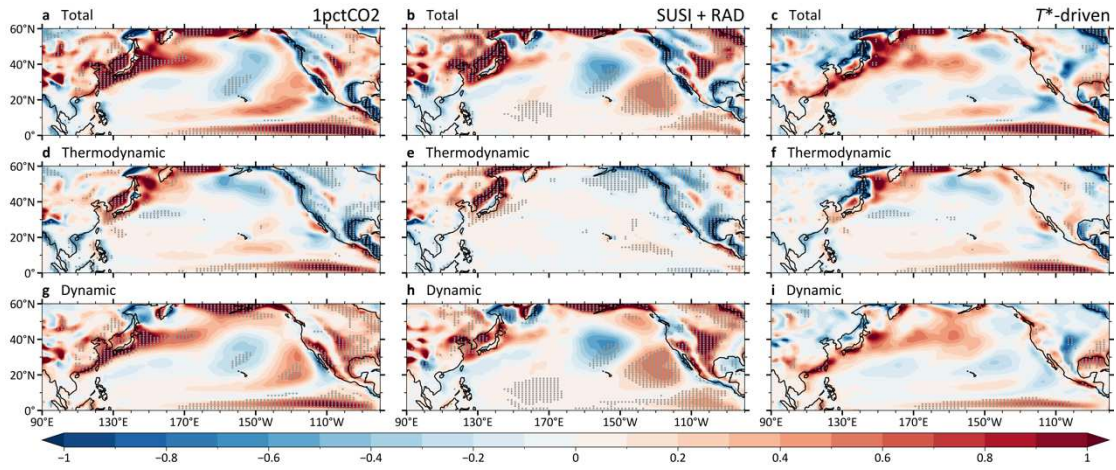
**Competing interests** The authors declare no competing interests.



**Extended Data Fig. 1** | Same as in **Fig. 1a**, but separated into (a) SUSI and (b) RAD. SUSI produces a significant SLP reduction across the Aleutian Low and NPSA, whereas an anomalous high near 50°N in RAD interrupts the connection between the two system changes (see Fig. 1a).



**Extended Data Fig. 2 | Heat budget decomposition (shading,  $\text{K K}^{-1}$ ) for diagnosing**  
**(see [Methods](#)) the SST warming pattern (contours, CI:  $0.1 \text{ K K}^{-1}$  below  $0.55 \text{ K K}^{-1}$**   
**and  $0.2 \text{ K K}^{-1}$  above  $0.75 \text{ K K}^{-1}$ ; red positive, blue negative; zero omitted) under**  
**the coupled 1pctCO2 scenario. By dividing the latent Newtonian plus longwave Plank**  
**cooling, the (a) total SST is decomposed into contributions from (b) radiative effect, (c)**  
**sensible surface stability, (d) latent surface stability, (e) sensible wind speed, (f) latent**  
**wind speed, (g) oceanic flux, and (h) latent relative humidity. Stippling denotes that the**  
**ensemble-mean SST component exceeds its intermodel spread, and a positive flux**  
**represents ocean warming. The strong hook string is dominated by surface stability and**  
**radiative effect, whereas the moderate tip is controlled by latent wind speed and relative**  
**humidity.**



**Extended Data Fig. 3 | Decomposition of SAT advection ( $10^{-6} \text{ K s}^{-1} \text{ K}^{-1}$ ) for**

**diagnosing (see [Methods](#)) the surface stability effect on the hook string.** Each row

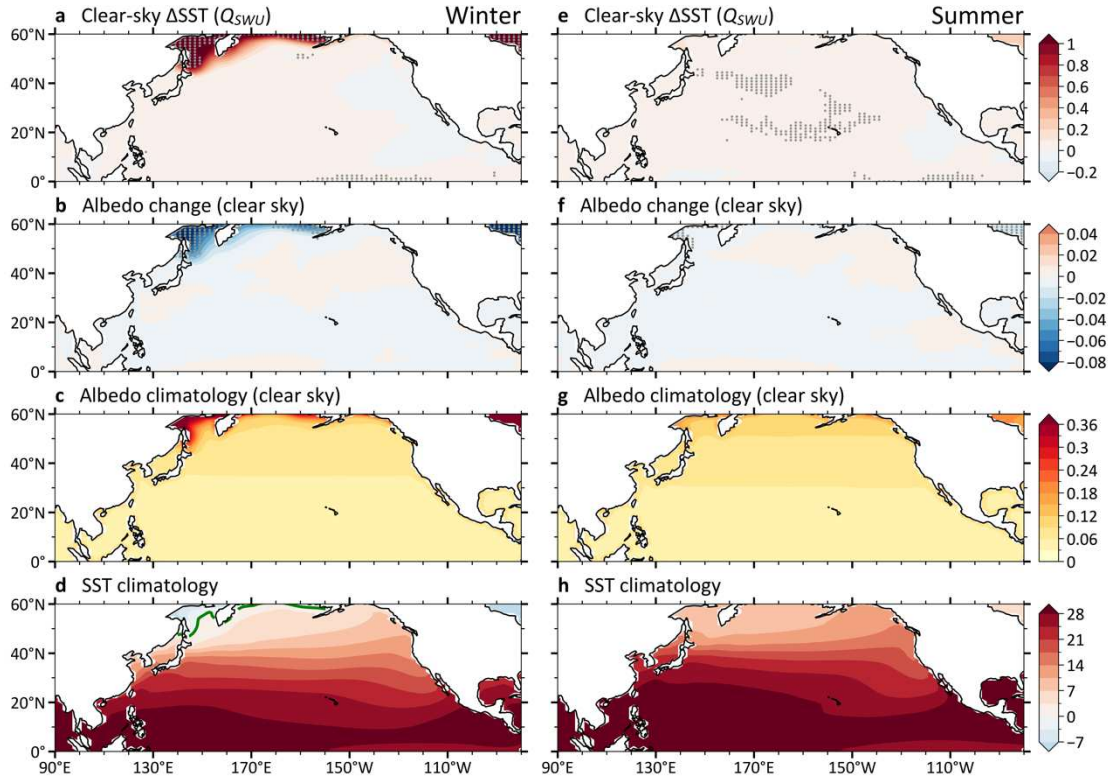
shows one component: the **(a-c)** total, **(d-f)** thermodynamic, and **(g-i)** dynamic

components – for the coupled 1pctCO2 projection (left), SUSI plus RAD (middle), and

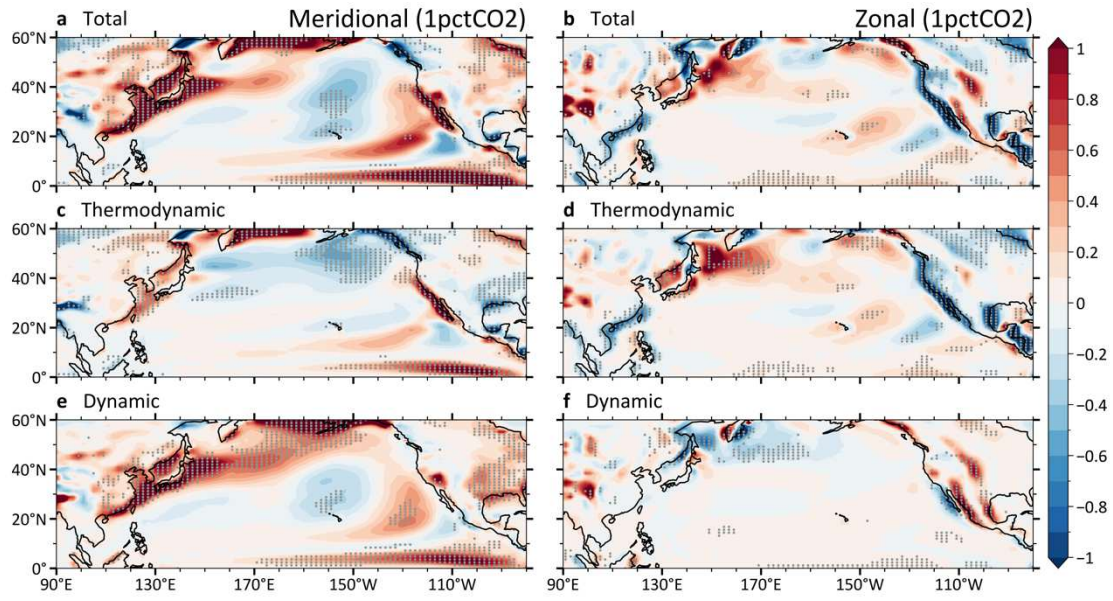
$T^*$ -driven effect (right). *The total advection matches the surface stability contribution*

*to the string warming, where the thermodynamic component dominates its northern*

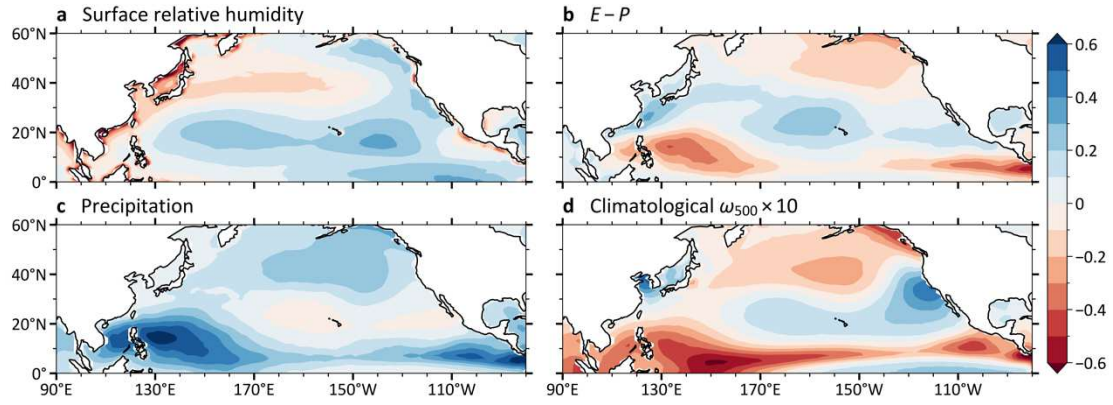
*portion and the dynamic its southern, so the string is jointly shaped by both pathways.*



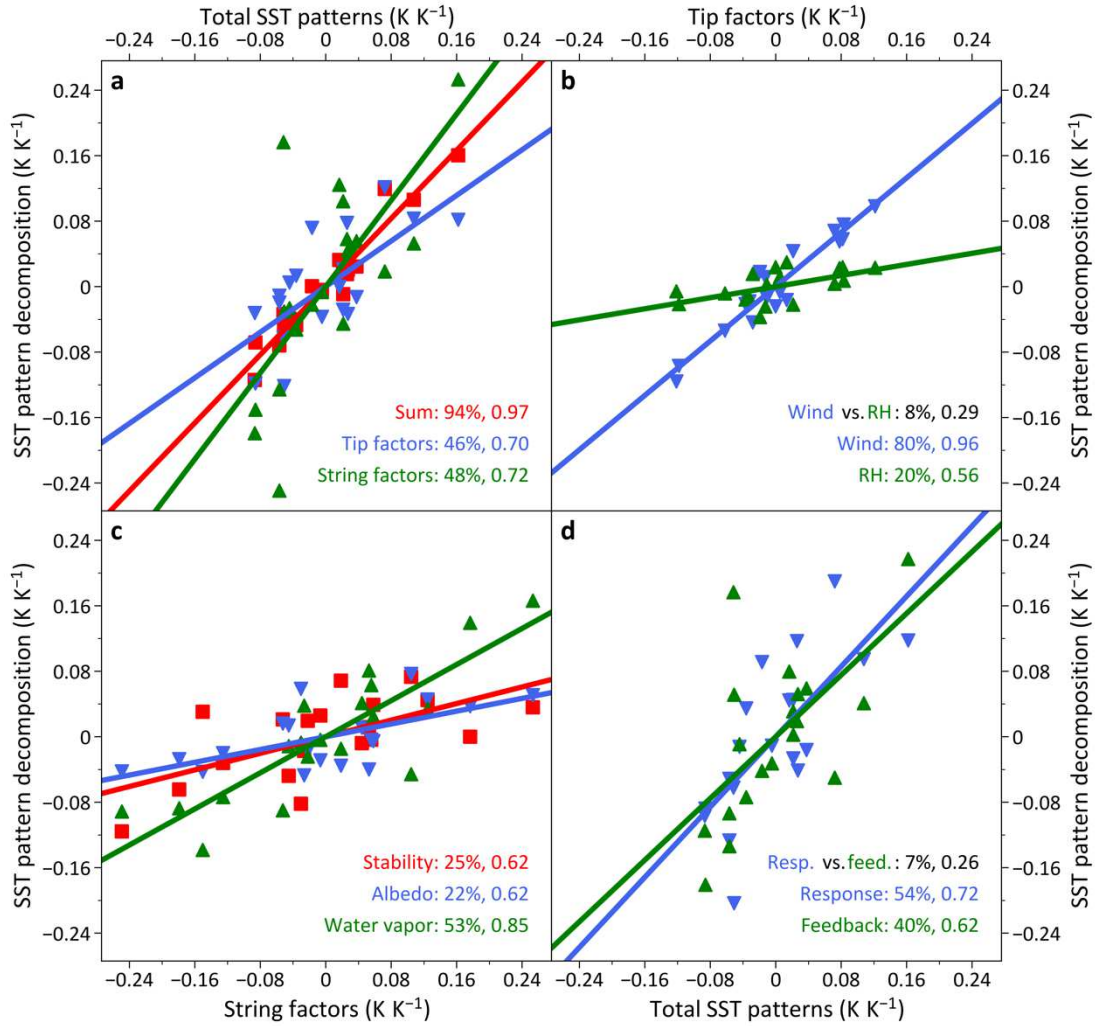
**Extended Data Fig. 4 | Clear-sky upward shortwave effect due to albedo feedback in the  $T^*$ -driven projection, for boreal (a-d) winter and (e-h) summer.** Changes in clear-sky upward shortwave effect (a, e)  $\Delta SST (Q_{SWU})$  ( $K K^{-1}$ ) is diagnosed with clear-sky albedo changes (fraction  $K^{-1}$ ) (b, f), climatological clear-sky albedo (fraction) (c, g), and climatological SST ( $^{\circ}C$ ) (d, h). The thick contour in (d) marks the  $0^{\circ}C$  isotherm. Stippling for the change fields indicates that the ensemble mean exceeds the intermodel spread. All panels are ensemble means over the 19 models providing clear-sky output. Winter is defined as January-June in (a-c) and December-May in (d) according to the climatological emergence of high albedo and  $0^{\circ}C$  SST, respectively. The one-month lag therefore represents the delayed sea-ice response to the SST seasonality. Summer covers the rest months in each case. Apparently, the clear-sky upward shortwave effect is attributed to winter albedo feedback by melting sea ice.



**Extended Data Fig. 5 | Same as the 1pctCO2 panels in Extended Data Fig. 3 for the SAT advection decomposition, but with each component separated into meridional (left) and zonal (right) directions. The thermodynamic component is projected mainly onto the zonal direction, while the dynamic component is projected mainly onto the meridional direction.**



**Extended Data Fig. 6 | Hydrological diagnostics in SUSI: changes in (a) surface relative humidity ( $\% \text{ K}^{-1}$ ), (b)  $E - P$  ( $\text{mm day}^{-1} \text{ K}^{-1}$ ), and (c) precipitation ( $\text{mm day}^{-1} \text{ K}^{-1}$ ), as well as (d) the climatological 500-hPa pressure velocity,  $\omega_{500}$  ( $\text{Pa s}^{-1}$ ). The tip lies in a subsidence region where relative humidity increases with enhanced  $E - P$  due to suppressed precipitation, indicating the dry-get-drier paradigm.**



**Extended Data Fig. 7 | Intermodel scatterplots of the regional-mean SST warming pattern ( $\text{K K}^{-1}$ ) under 1pctCO2 (Extended Data Fig. 2) against the key factors in the heat budget.** The hook string and tip regions are defined as  $25^{\circ}$ - $60^{\circ}\text{N}$  and  $5^{\circ}$ - $25^{\circ}\text{N}$ , respectively, over  $130^{\circ}\text{E}$ - $110^{\circ}\text{W}$ . (a) The SST pattern regressed on the string and tip contributions and on the sum over  $5^{\circ}$ - $60^{\circ}\text{N}$ ; (b) the tip contribution regressed on latent wind speed and relative humidity; (c) the string contribution regressed on surface stability and on the albedo and water vapor feedbacks; and (d) the total regressed on the atmospheric responses (surface stability plus latent wind speed and relative humidity) and the albedo plus water vapor feedback. Each symbol is one model's intermodel anomaly. All panels use the same 19 models providing clear-sky output, so that the contributions remain directly comparable. The fractional contribution of each factor is computed with the novel  $R^2$  method (see Methods). The sum reconstruct the total, with the tip dominated by wind speed and the string shaped by both stability response and radiative feedbacks; overall, the atmospheric responses and albedo plus water vapor feedback contribute comparably to the total pattern.

## Supplementary Files

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