



A method for seeding and unseeding tropical cyclones in Earth system models using axisymmetric vortices

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Abstract. The upscale influence of tropical cyclones (TCs) on the broader climate system remains poorly understood because of limitations in existing diagnostic and modeling methods. To address this, we present a method to modulate TC activity online in Earth-system models. Balanced, axisymmetric TC-like vortical perturbations derived from an analytical formulation with tunable parameters are superimposed on the atmospheric state. To add TCs to the model (“seeding”), cyclonic perturbations are introduced into quiescent environments. To remove TCs from the model (“unseeding”), anticyclonic perturbations are iteratively tuned to cancel existing TCs. During vortex insertion and removal, the balanced formulation prevents spurious wave generation while scalar global field corrections prevent climate drift.

In fully coupled aquaplanet simulations, the method reliably modulates TC climatology: seeding increases TC frequency and accumulated cyclone energy well above baseline, while unseeding reduces them by up to 95%. TC tracking shows that seeding primarily enhances genesis, whereas unseeding suppresses TC frequency by predominantly targeting incipient storms. The resulting experiments produce systematic tropical mean-state and radiative responses that scale approximately monotonically with TC activity, including upper-ocean cooling and mid-tropospheric drying as TC activity increases. These large-scale atmospheric mean-state responses differ qualitatively from short-term local anomalies associated with individual TCs, providing strong evidence that TCs modify their surrounding environment. These results establish online vortex manipulation as a practical, climatologically stable framework for investigations of TC upscale effects in the coupled tropical atmosphere-ocean system.

1 Introduction

The climate and extreme weather research communities have extensively studied the influence of the climate on tropical cyclone (TC) genesis and intensity. Empirical and theoretical frameworks that explain environmental control of TCs have been developed and refined, accounting for influences from sea-surface temperatures (SSTs), the vertical temperature profile of the atmosphere, water vapor, environmental wind shear, and other factors (Gray, 1979; Emanuel, 1986; Emanuel et al., 2004; Kieu and Wang, 2017). These frameworks have been applied to predict how forced climate change and slower natural modes of variability modulate TC activity. On the other hand, the upscale effects of TCs on the broader climate system (i.e., how TCs affect circulations at larger spatial scales and longer timescales than TCs themselves) are poorly understood even though TCs have long been known to influence their surrounding environment (Fisher, 1958; Gangopadhyaya and Riehl, 1959).



Understanding these effects may provide avenues for improving subseasonal-to-seasonal prediction. At longer timescales, climate change will alter the intensity and frequency of TCs along with the rainfall they produce (Knutson et al., 2020). The resulting upscale effects may become a critical factor in climate feedbacks that modulate the overall climate response to anthropogenic emissions. Additionally, general circulation models (GCMs) are known to have errors in their representation of TCs due to insufficient atmospheric resolution (Bengtsson et al., 1995; Camargo and Wing, 2016) and simplified ocean coupling (Zarzycki, 2016). Such errors may lead to deficiencies in simulations of the Earth system, underscoring the need to clarify TC upscale effects.

The upper ocean mediates a well-established pathway for some of these upscale interactions. For several decades, observational evidence (Fisher, 1958; McFadden, 1967) and modeling (Chang and Anthes, 1978; Price, 1981) have consistently shown that the sea surface cools during the TC passage. The cold SST anomalies, or “cold wakes,” persist for up to one month, generating anomalous surface heat fluxes (reduced upwelling longwave, sensible, and latent heat fluxes) as SSTs recover to climatology (Dare and McBride, 2011). These anomalous fluxes contribute greatly to the annual mean ocean heat uptake (OHU) (Emanuel, 2001; Jansen et al., 2010; Buetti et al., 2014) and can also feed back on the thermodynamic state of the atmosphere (Hart et al., 2007; Sobel and Camargo, 2005; Schenkel and Hart, 2015b). As higher resolution climate models allow for more accurate simulations of TC intensity, it is becoming clearer that TC wind forcing on the ocean surface mediates a substantial upscale influence of TCs on the climate (Sriver and Huber, 2007; Li et al., 2023). Similarly, finer ocean model grids may highlight the role of small-scale ocean processes in modulating TC-induced ocean heating (Brizuela et al., 2023).

The direct upscale influence of TCs on the atmosphere, bypassing communication through ocean heat anomalies, is less clear. From a purely linear additive viewpoint, the localized circulation and state anomalies associated with TCs significantly alter the long-term mean and variability of the atmosphere (Hsu et al., 2008; Arakane and Hsu, 2021). Arakane and Hsu (2021) coined the term “TC footprints” to describe the difference between the statistics of reanalysis state fields with TCs removed and the statistics of the unmodified reanalysis. They found that TC footprints contribute substantially to the long-term circulation, but these footprints cannot account for nonlinear TC-climate interactions involving other modes of variability. For instance, TCs are known to dynamically interact with the large-scale environment by exciting Rossby waves (Sardeshmukh and Hoskins, 1988; McDonald, 1998; Schenkel and Hart, 2015b) and by modifying extratropical wave packets through potential vorticity advection (Hodyss and Hendricks, 2010; Archambault et al., 2013).

TC interactions with hydroclimate extremes and the hydrological cycle have been studied extensively and likely also account for a major upscale influence on the tropical climate. Regionally within active basins, TCs account for a major fraction of annual total precipitation (Konrad and Perry, 2010; Hsu et al., 2008) and horizontal moisture transports (Xu et al., 2017; Guo et al., 2017; Liu et al., 2021; Pérez-Alarcón et al., 2023). Through these anomalous moisture fluxes, TCs can induce synoptic- and basin-scale anomalies in water vapor content (Scoccimarro et al., 2020; Schenkel and Hart, 2015a). Though the earlier studies have established mature TCs to be highly effective moisture transporters, more recent theoretical work posits that developing TCs may fundamentally modulate the mean tropical circulation by exporting water vapor poleward (Adames and Ming, 2018; Adames Corraliza and Mayta, 2024; Lin et al., 2025).



60 Although these all constitute plausible mechanisms by which TCs interact with larger-scale atmospheric circulations, re-
 search has been relatively scant regarding their comparative influences on the long-term climate state and general circulation.
 Recently, Rios et al. (2026) provided a major advancement by globally reducing TC activity in a coupled climate model,
 demonstrating that TCs drive large-scale radiative cooling by drying the free troposphere and reducing high cloud coverage.
 However, while their work establishes the net macroscopic climate response to TCs, isolating and directly comparing the
 65 individual atmospheric pathways that drive these changes remains a major challenge.

To complement the experimental techniques of Rios et al. (2026) and help establish causal inference in studying TC upscale
 effects, we employ a novel technique to artificially modulate TCs. Past modeling studies have reduced TC intensity by denying
 the wind-induced surface heat exchange (WISHE) intensification mechanism of TCs (Montgomery et al., 2009; Zhang and
 Emanuel, 2016; Ma and Fei, 2022), but these studies have largely focused on storm dynamics in limited-area models without
 70 addressing the upscale effects of TCs on the general circulation. Rios et al. (2025) devised a version of this artificial evaporative
 suppression technique known as “SWISHE” (Suppression of WISHE), where parameterized surface latent heat fluxes are
 adjusted downward within TC grid cells. Applying SWISHE globally in three different Earth-like GCMs, they obtained a 44%
 decrease in the number of TCs and a 90% decrease in hurricane-strength TCs.

An alternative class of methods for artificially controlling TCs online in numerical models is through vortex insertion (bo-
 75 gusing) or removal. These vortex manipulation methods have typically been used in weather forecasting contexts to force TC
 initialization in lower-resolution numerical weather prediction or to assess the sensitivity of TC evolution to initial conditions
 (Leslie and Holland, 1995; Kurihara et al., 1990; Reed and Jablonowski, 2011). Just as no study had employed WISHE denial
 in Earth-system modeling experiments prior to Rios et al. (2025), our search revealed no prior use of TC vortex insertion or re-
 moval to examine TC upscale effects. Although vortex manipulation is established in regional or short-term numerical weather
 80 prediction, manipulating TC vortices in climate model experiments requires extra care to maintain local dynamical balance and
 conserve global energy, mass, and momentum to prevent climatological drift. If vortex manipulation can be adapted to these
 constraints, it offers some desirable properties. First, vortex manipulation can allow control over the number of TCs rather
 than just their intensity, though the resulting TC climatology remains an emergent property of the model background state
 (Sect. 3.1). Second, vortex manipulation entails a model state modification rather than altering the underlying (surface heat
 85 flux) physics of the model. This ensures that the air-sea thermodynamic coupling mechanisms by which TCs induce upscale
 influences remain physically consistent. Changes in the climate statistics can then be attributed to changes in dynamically
 coherent structures (TC) rather than perturbations to underlying physics. Third, and central to this paper, strictly adhering to
 local balance and global conservation constraints enables us to confidently attribute changes in climate statistics to the altered
 TC climatology rather than artifacts of vortex insertion and removal.

90 In this paper, we develop and evaluate an online vortex-manipulation technique that can effectively control the TC activity
 in an Earth-like aquaplanet while satisfying these constraints. Sect. 2 describes the seeding and unseeding methods and the
 associated TC tracking. Sect. 3.1 evaluates how these interventions alter TC frequency, intensity, and genesis. Sects. 3.2–
 3.3 then document the resulting tropical thermodynamic, radiative, and circulation responses and compare them with the



direct footprints of individual TCs. Together, these analyses establish the method and identify large-scale climate adjustments
 95 requiring more detailed process-level investigation.

2 Methods

2.1 Idealized models and the role of AQUA

Comprehensive Earth-system models (ESMs) can reproduce a wide array of variability and feedbacks that are present in the real Earth system. Although they simulate the climate of Earth realistically, their sophisticated dynamical behavior can make
 100 it difficult to diagnose what basic processes give rise to observed phenomena. To simplify model behavior, researchers have created a hierarchy of climate models over time, ranging from two-layer quasigeostrophic models to dry atmospheric GCMs to comprehensive climate models (Held, 2005). Within this hierarchy lie aquaplanet models, which simplify the lower boundary condition of the atmosphere by removing landmasses and orography (Neale and Hoskins, 2000; Medeiros et al., 2016; Wu et al., 2021). Depending on the configuration, aquaplanet models can simulate the mean meridional circulation (MMC) of
 105 the atmosphere (Neale and Hoskins, 2000), the Intertropical Convergence Zone (ITCZ; Goswami et al., 1984), variability of atmospheric jets (Marshall et al., 2007), and the control of ocean circulations on the meridional heat transport of the ocean (Smith et al., 2006). Idealized studies of TCs have also used aquaplanets to understand environmental controls on TCs, reducing our reliance on empirical approaches (Merlis and Held, 2019). The aquaplanet model described by Wu et al. (2021) includes a fully coupled ocean model and the full physical parameterizations of more comprehensive atmosphere-ocean GCMs. This
 110 class of aquaplanet models offers an opportunity to explore the upscale effects of TCs on the dominant meridional circulations and heat transports in the global atmosphere and ocean.

We run a fully coupled aquaplanet configuration (Youwei Ma, through personal communication) of the Community Earth System Model (CESM) based on the “Aqua” configuration described by Wu et al. (2021). For the rest of this paper, we refer to our aquaplanet configuration as “AQUA” even though model components and forcing fields may differ slightly from
 115 those of Wu et al. (2021). In our configuration, the atmosphere is modeled by the Community Atmosphere Model (CAM) spectral element dynamical core (CAM-SE; Lauritzen et al., 2018) with version 4 physics (CAM4; Neale et al., 2010). Both the dynamical core and physics run on a common 0.25° (~ 28 km) cubed-sphere grid with 120 cells along each edge of the cube and four Gauss-Lobatto nodes per cell in each direction ($n_e = 120, n_p = 4$ or $ne120np4$). The Zhang-McFarlane deep convective scheme (Zhang and McFarlane, 1995) is used along with the Hack parameterization of shallow convection (Hack,
 120 1994). To simulate ocean dynamics, the Modular Ocean Model version 6 (MOM6) is run with a uniform 4-km ocean depth over the domain 80°S – 80°N . Land caps cover the regions poleward of 80° latitude. The sea ice model is dynamic-thermodynamic, and there is zero heat flux at the ocean floor. The ocean zonal grid spacing is uniform at $2/3^\circ$ (~ 74 km), and the meridional grid spacing smoothly varies from $2/3^\circ$ at 20° latitude to $1/8^\circ$ at the poles. Refinement of the meridional grid spacing down to $1/4^\circ$ is used within 20° of the equator to resolve equatorial upwelling. Mesoscale eddy transports are represented by the
 125 Gent-McWilliams scheme (Gent and McWilliams, 1990). Apart from the diurnal cycle, there are no zonal asymmetries in the boundary conditions or forcing of the aquaplanet. A realistic seasonal cycle in insolation is also imposed, with the vernal



equinox fixed on 21 March. The seasonal cycle induces seasonal deepening and shoaling of the ocean mixed layer, which plays a vital role in modulating TC-induced ocean heat uptake (Stommel, 1979; Jansen et al., 2010). The boundary conditions of AQUA are hemispherically symmetric, and the annual-mean insolation forcing is hemispherically symmetric.

130 To equilibrate the deep ocean to the forcings used in our AQUA configuration, the model was spun up in several stages. First, the ocean state was initialized from Year 410 of the simulation described by Wu et al. (2021) (Youwei Ma, personal communication), and the model was run freely for a further 1100 years with CAM's finite-volume atmospheric dynamical core on a $0.9^\circ \times 1.25^\circ$ latitude-longitude grid (for computational expediency), at which point the top-of-model radiative balance was $+1.5 \text{ W m}^{-2}$ (net warming). Integration was resumed from the same ocean state for 30 years with a $\sim 1^\circ$ cubed-sphere spectral
 135 element atmospheric grid (ne30np4). Afterwards, the relative humidity (RH) thresholds for cloud formation were adjusted in the cloud microphysics scheme to 80% for high clouds and 89% for low clouds to achieve top-of-model radiative balance during the following 20 years. Next, the atmospheric grid was switched to ne120np4 while retaining the ocean state, and we applied two additional physics modifications: the ice-cloud modification of Bacmeister et al. (2014) and a reduction in the Zhang-McFarlane convective adjustment timescale from 3600 s to 1800 s. This high-resolution run was integrated forward for
 140 four years with a high-cloud RH threshold of 80% and a low-cloud RH threshold of 88.5%, achieving a radiative imbalance of $+0.1 \text{ W m}^{-2}$ and arriving at the initial conditions for the model experiments in this paper. The initial conditions contain a hemispheric asymmetry, with permanent Antarctic ice and only seasonal Arctic ice; this asymmetry persists across experiments but does not materially affect the tropical or subtropical model state and is noted here as a consequence of the coupled-model spinup. The timestamp of these initial conditions is referred to as 1 February of Year 0005 in the text and figures.

145 2.2 TC tracking

TCs are tracked on the native atmospheric grid using TempestExtremes (Ullrich and Zarzycki, 2017). The `DetectNodes` binary is used for both online TC detection (as part of the seeding and unseeding process) and offline TC tracking. The `DetectNodes` call closely follows that in Section 3.2 of Ullrich et al. (2021). The algorithm searches for local minima in SLP that are at least `PSLFOMAG` lower than the pressure at distance `PSLFODIST` away. The node location is defined by
 150 the center of the mesh face containing the SLP minimum. Additional filtering steps are applied using a local 300–500 hPa thickness maximum (to enforce an upper-tropospheric warm core) and, in some cases, an 850 hPa cyclonic relative vorticity maximum. The local extrema of these additional filtering variables must fall within the distances `WCOFFSET` and `RVOFFSET`, respectively, from the sea-level pressure minimum to exclude excessively sheared disturbances. `DetectNodes` thresholds can be found in Appendix A.

155 For online TC detection during seeding and unseeding, only a single model snapshot is used to locate TC-like nodal features, so only the `DetectNodes` routine is called. A key distinction between the online and offline criteria is that `WCOFFSET` is restricted to 0.5° great-circle distance (GCD) in the former, compared to 1° GCD in the latter. This stricter online criterion is required to filter out vertically tilted systems and isolate only highly upright, axisymmetric cyclones for real-time modification. During offline tracking, `StitchNodes` is used as in Ullrich et al. (2021) to assemble the nodal features found by
 160 `DetectNodes` into TC trajectories. `StitchNodes` thresholds ensure that storms can be continuously tracked for a sufficient



duration with a minimum wind speed. The corresponding thresholds, `TRAJMINLENGTH` and `MINWIND`, and other criteria can be found in Appendix A.

Throughout this paper, we use the term “candidate vortices” to refer to nodal features that are considered in the online unseeding and seeding process, identified by the online `DetectNodes` configuration. We reserve the term “discrete TCs” (or simply “TCs”) for storms that satisfy the full offline tracking criteria, including trajectory stitching by `StitchNodes`. These categories heavily overlap but are not identical: only some candidate vortices represent a snapshot of a discrete TC while others dissipate before satisfying offline tracking criteria. Conversely, some discrete TCs may not be identified as candidate vortices during seeding and unseeding because of stricter online detection thresholds.

2.3 Seeding and unseeding TCs

Here, TCs can be either artificially added to or removed from the ambient flow. While we acknowledge some ambiguity in the term “TC seed” (e.g., Wang et al., 2025), we use “seeding” to refer to the artificial addition of TCs and “unseeding” to refer to the artificial removal of TCs. To seed and unseed TCs, the aquaplanet model is paused at a specified interval, and a snapshot of the atmospheric state is saved as a restart file. The `DetectNodes` binary in `TempestExtremes` (Ullrich and Zarzycki, 2017) is applied to these snapshots to locate candidate vortices as defined in Sect. 2.2. In both seeding and unseeding, the model restart file is then directly modified, and the model is subsequently reinitialized and freely integrated forward until the next un/seeding time. Importantly, when no state modification is applied, this framework preserves bit-for-bit reproducibility with an uninterrupted integration. This approach allows for experimental flexibility, including control over the seeding rate (i.e., seeds per day), spatial location (both latitude and longitude), and proximity between seeds and existing candidate vortices.

We seed and unseed TCs by leveraging the work of Reed and Jablonowski (2011), who presented an analytic method of initializing TCs in numerical models. Reed and Jablonowski derived analytical formulas for three-dimensional axisymmetric vortices. The analytical formulas provide the pressure and temperature profiles of warm-cored cyclonic vortices as perturbations that are embedded within a realistic tropical background state. The perturbations satisfy hydrostatic balance, and a tangential wind field is derived to satisfy gradient-wind balance. The balanced state of these so-called “Reed-Jablonowski vortices” (RJVs) prevents the formation of spurious inertia-gravity waves that would otherwise arise from inaccurate modification of the state fields in a numerical model. Parameters can be adjusted to control the minimum surface pressure deficit (dp), radial scale of the vortex (r_p), and the height of the warm core (z_p). Although the original formulation describes warm-cored cyclonic vortices used for seeding, we also invert the signs of RJV fields to construct cold-cored anticyclonic perturbations capable of canceling existing candidate vortices during unseeding. Throughout this paper, we use the term “intervention” to refer to any single, discrete instance where a RJV (either cyclonic or inverted) is superimposed on the model state.

To superimpose RJVs on the atmospheric state, we adjust the surface pressure, wind, and temperature fields. After inserting each vortex, a global correction is applied to preserve atmospheric mass and mass-weighted state field integrals. To preserve these quantities, we determine the residual difference between global integrals of the initial and modified states. These residuals are then zeroed by applying a uniform scalar correction throughout the global domain for each variable (p_s , T , q , u , v) in the modified state.



195 Seeding a TC involves superimposing an RJV on the existing atmospheric state fields. Canonically, weak TC precursor “seeds,” such as African Easterly Waves, are used in transition-probability frameworks (e.g., Emanuel, 2022). In this study, the inserted RJVs contain coherent TC-like circulations to improve their likelihood of survival and development into self-sustaining TCs. To define each seeded vortex, parameters controlling its amplitude, spatial structure, and location are randomly sampled from prescribed distributions. The specific distributions from which these parameters are drawn, whether uniform or derived
 200 from prior model results, are detailed for each experiment in Sect. 2.4. To prevent storms from overlapping, a minimum separation of 15° GCD is enforced between seeds and existing TC centers. Unlike the forthcoming unseeding protocol, the modification of the water vapor field during seeding is dependent on the specific experimental design, as detailed in Sect. 2.4.

Two examples of state modifications to seed Northern Hemisphere TCs in SEED are shown as constant-latitude cross sections in Figure 1. The original quiescent states (left column) display minimal lateral gradients in θ_e and the meridional wind
 205 fields. To seed TCs, RJV perturbations (center column) are imposed with a warm core (solid T' contours), a low-pressure center (dashed p' contours), cyclonic winds (southerly winds located west of northerly winds), and additional water vapor (green shading of δq). The modified states (right column) show the RJVs superimposed on the originally quiescent states, where they either sustain themselves during subsequent model integration or dissipate in unfavorable environments.

Unseeding entails removing existing candidate vortices from a given atmospheric state by applying an inverted (i.e., anti-
 210 cyclonic) RJV directly over the identified low-pressure center. Rather than using static vortex parameters to unseed a diverse population of storms, we employ a novel optimization routine that fits the analytical RJV to each candidate vortex. This routine iteratively adjusts the three-dimensional RJV parameters to minimize the root-mean-square error between the analytical and modeled pressure and temperature anomalies. To isolate the candidate vortex structure for this fit, the surface pressure and 3D temperature profiles are estimated by azimuthally averaging the model fields out to a radius of 800 km from the SLP minimum.
 215 The candidate vortex profile is then expressed as anomalies relative to this environmental boundary before the fit is performed. Though this approach lacks the precision of vortex removal techniques that use potential vorticity (PV) inversion, it has a lower computational cost and better numerical stability. For the water vapor field, a decrement in specific humidity q is first calculated based on the temperature adjustment such that the relative humidity would remain constant in every grid cell. The decrement in q is then multiplied by a constant factor of 2.5 (“ q factor” in Table 1) and applied to the model state. This constant factor was
 220 determined empirically to remove local specific humidity anomalies and prevent frequent TC (re)development after unseeding. A typical unseeding event removes $\mathcal{O}(10^{12})$ kg of water vapor locally; the subsequent global scalar correction restores global water vapor content to within $\mathcal{O}(10$ kg) or $\mathcal{O}(10^{-13} \%)$ of its original value.

Two examples of state modifications to remove Northern Hemisphere TCs from UNSEED are shown as constant-latitude cross sections in Fig. 2. The original states (left column) contain TC-like candidate vortices characterized by warm-core, low-
 225 pressure systems with cyclonic winds and moist centers. To cancel out these candidate vortices, the imposed perturbation (center column) consists of a temperature decrease (dashed T' contours), a pressure increase (solid p' contours), anticyclonic winds (represented by southerly flow to the west of the center and northerly flow to the east of the center), and moisture removal (brown shading). In the modified states (right column), the pressure anomaly field appears uniform, and the T' and θ_e fields no longer show marked warm and moist anomalies. In both cases, the original v -winds were asymmetric about the candidate

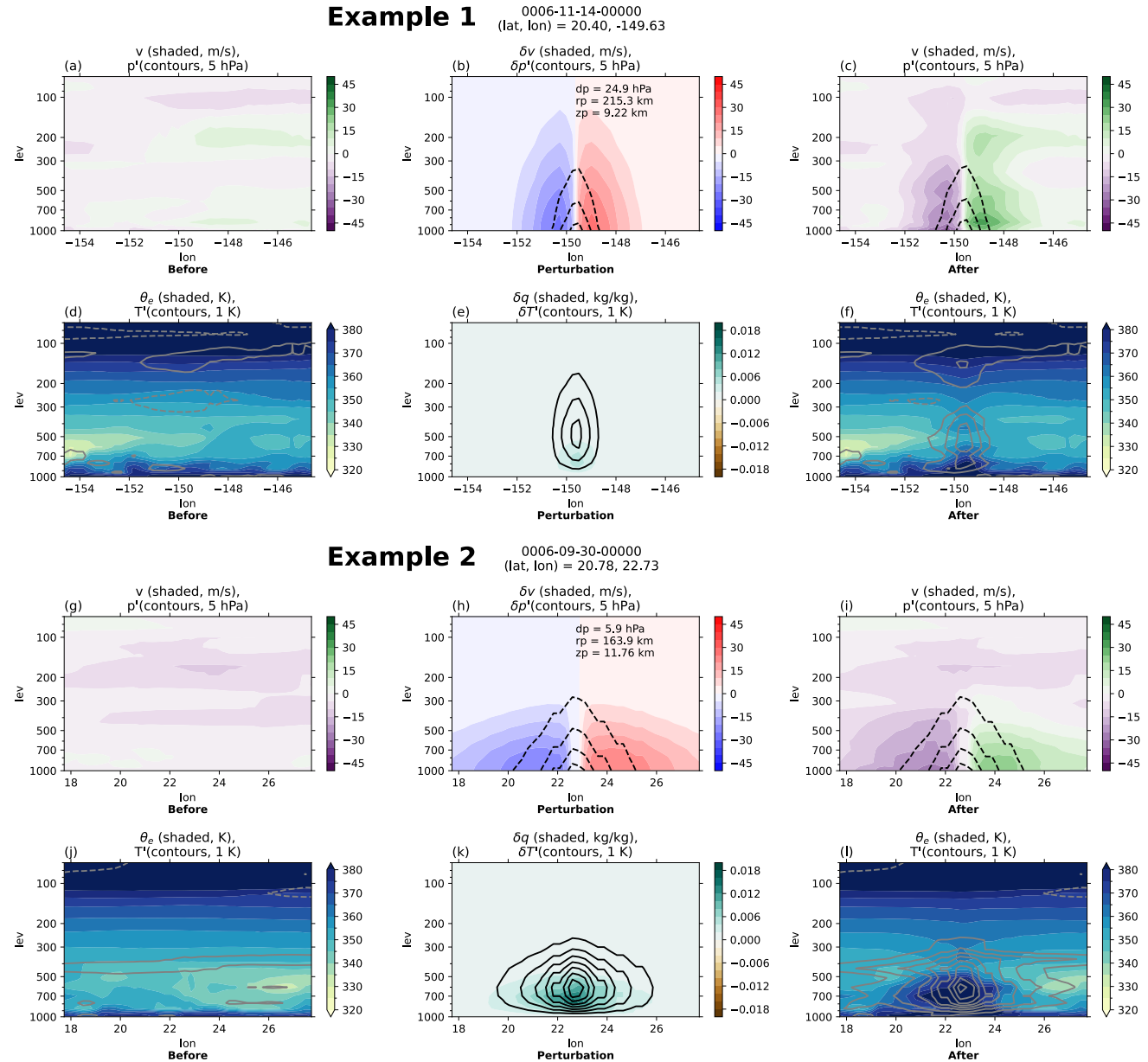


Figure 1. Constant-latitude cross sections of two distinct seeding interventions, labeled Examples 1 and 2, and displayed in panels (a)–(f) and (g)–(l) on native model levels. The three columns show the unmodified model state prior to seeding (left), the idealized RJV perturbation (middle), and the modified model state with the perturbation superimposed (right). For each seeding event, the upper row displays the meridional wind component v (shaded) and pressure anomaly p' (contoured). The lower row displays the equivalent potential temperature θ_e (shaded) for the “Before” and “After” states, the specific humidity perturbation δq (shaded) for the middle column, and the temperature anomaly T' (contoured) across all panels. Anomalies denoted with a prime are evaluated relative to the azimuthal mean value 5 great-circle degrees from the center of the storm.



230 vortex centers, resulting in a modified state with residual vorticity. Despite the imposed axisymmetric perturbations lacking the precision of potential vorticity inversion vortex removal techniques, the modified states confirm a near-total cancellation of the low-pressure center and the warm core.

To verify that these state modifications do not provoke numerical shock or generate spurious inertia-gravity waves during forward integration, Fig. 3 illustrates the short-term prognostic evolution of the sea-level pressure (SLP) field using the same example cases featured in Figs. 1–2. The top row tracking the unseeding event (Example 1 from Fig. 2) shows a well-defined SLP minimum (dark purple circle) six hours before intervention (Fig. 3a). Because model history files are output immediately prior to restart modification, the hour-0 snapshot still reflects this candidate vortex (Fig. 3b). In the 6 to 12 hours following the inverted vortex superposition, the local SLP minimum vanishes from an otherwise quiescent background environment without radiating grid-scale wave artifacts (Fig. 3c,d). Conversely, the seeding sequence in the bottom row (Example 2 from Fig. 1) demonstrates an analogous, stable progression in reverse: inserting a balanced RJV into a quiet environment yields a coherent cyclonic disturbance that evolves cleanly over the subsequent integration hours.

2.4 Model experiments

We conducted five model experiments in AQUA, differing only in artificial TC manipulation (Table 1). A free-running control experiment (CTL) serves as a baseline with no TC manipulation. In all other experiments, TCs are only manipulated in the hemisphere of maximum zonal-mean SST, approximating the summer and fall seasons. We chose this hemispheric restriction to leave intact any interhemispheric or interseasonal climate responses that may result from TCs. While the average dates when the SST maximum switches hemispheres differ between experiments (Fig. 4), we standardize the definition of “warm season” to cover Northern Hemisphere June–November and Southern Hemisphere December–May for all experiments. The treatment of seasons is further detailed in Sect. 3.1.

250 The perturbed model experiments, UNSEED_EX, UNSEED, SEED, and SEED_EX, are named according to the type of TC modulation imposed (unseeding or seeding), where the _EX suffix denotes a more extreme imposed forcing compared to experiments without the suffix. In UNSEED runs, the model is paused every 24 hours, and all detected candidate vortices in the hemisphere containing the maximum zonal-mean SST are unseeded as described in Sect. 2.3. To avoid excising gridpoint storms and convective features other than TCs in UNSEED, we also enforced a cyclonic vorticity threshold when selecting candidates for removal. On the other hand, UNSEED_EX was intended to eliminate as many TCs as possible, and thus was run with no vorticity threshold and more permissive pressure/warm-core thresholds. Although it would be ideal to perform unseeding at every dynamical timestep, the computational overhead of model restarts makes it impractical to further reduce the 24-hour restart interval without modifying low-level CESM source code.

For seeding experiments, the model is also paused every 24 hours to allow vortex insertion. In SEED_EX, a single RJV is seeded at every restart using parameters sampled from uniform distributions (15–40 hPa pressure deficit and 150–450 km RMW). The high intensity of seeds produces an increase in TC activity in SEED_EX that is several times the CTL baseline (Table 1). The moisture field is not modified in SEED_EX as most seeds become self-sustaining without artificial moisture injection due to their high amplitude at insertion. For a more moderate forcing experiment, the forcing imposed in SEED

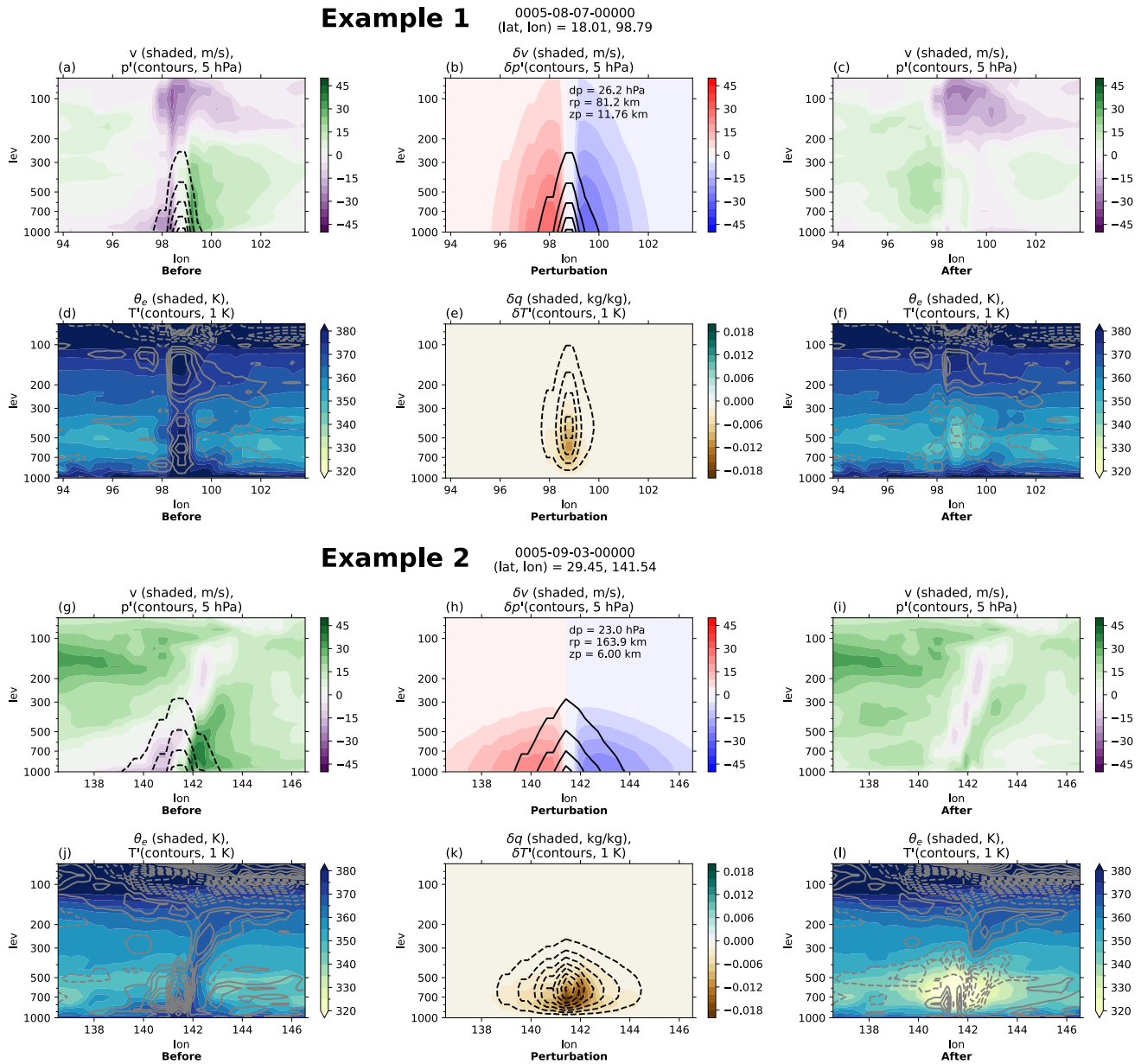


Figure 2. Constant-latitude cross sections of two distinct unseeding interventions, displayed in panels (a)–(f) and (g)–(l). The three columns show the unmodified model state with mature TCs prior to unseeding (left), the inverted RJV perturbation (middle), and the modified model state with the perturbation superimposed (right). Variables plotted are as in Fig. 1

Table 1. Description and vortex perturbation parameters of each experiment.

	UNSEED_EX	UNSEED	CTL	SEED	SEED_EX
Purpose	Aggressive seeding to minimize TC activity	Moderate seeding	Obtain a baseline climatology	Mirror the forcing in UNSEED	Aggressive seeding to probe upper bound
Description	Unseed all TCs in warm season DetectNodes thresholds SLP: 4 hPa $\rightarrow$ 2 hPa, DZ: 15 m $\rightarrow$ 6 m	Unseed all TCs in warm season Default online DetectNodes thresholds plus ζ threshold: $8 \times 10^{-5} \text{ s}^{-1}$	Free-running	Annual count and vortex parameters matched to UNSEED	1 seed per day in warm season
Intervention interval	24 hours	24 hours	—	24 hours	24 hours
Annual intervention count	647.8	212.3	—	213.9	364.7
Parameter sampling distributions	Natural (DetectNodes)	Natural (DetectNodes)	—	Matched to UNSEED	Lat: $\mathcal{U}(5^\circ, 20^\circ)$, dp : $\mathcal{U}(15 \text{ hPa}, 40 \text{ hPa})$, RMW: $\mathcal{U}(150 \text{ km}, 450 \text{ km})$
dp [hPa]	4.82 (2.10, 9.35)	9.05 (4.68, 18.12)	—	9.14 (4.89, 18.33)	27.45 (16.27, 38.66)
r_p [km]	210.9 (73.3, 306.7)	188.8 (73.3, 306.7)	—	189.0 (73.3, 306.7)	341.3 (176.9, 515.0)
$ \phi_c $ [°]	10.9 (5.9, 20.2)	12.4 (6.4, 22.3)	—	16.3 (8.2, 24.3)	12.1 (5.6, 19.2)
q factor	2.5	2.5	—	2.5	0

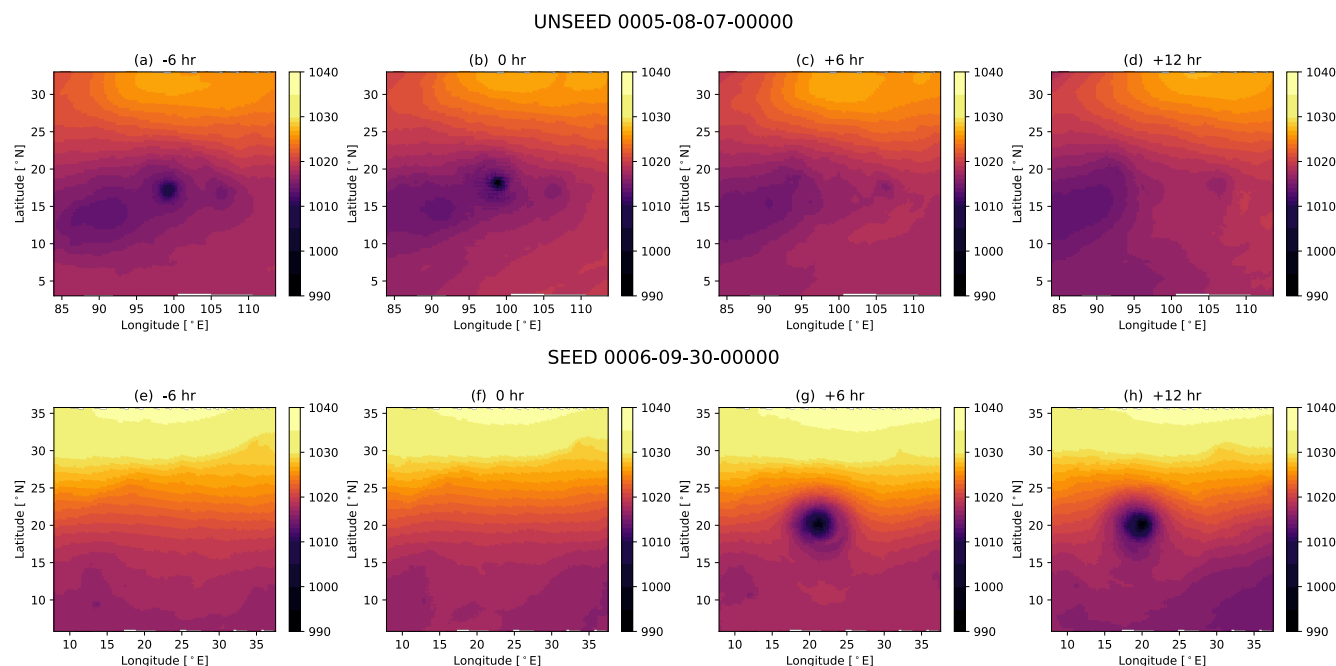


Figure 3. Prognostic evolution of the sea-level pressure (SLP; shaded [hPa]) field surrounding the time of vortex insertion/removal for (top) UNSEED Example 1 from Fig. 2, and (bottom) SEED Example 2 from Fig. 1. Hour 0 denotes the model state immediately prior to restart modification ($t = 0$ h). The SLP contour interval is set to 1 hPa over the range 1010–1030 hPa to highlight the lack of grid-scale wave artifacts at hours +6 and +12 following the superposition of balanced RJVs.

is designed to mirror that of UNSEED. First, the probability of seeding a vortex at each 24-hour restart is tuned to 58.1%,
 265 yielding the same mean intervention frequency (212 per year) as UNSEED. Second, rather than drawing RJV parameters
 from uniform distributions, the dp , r_p , and central coordinates are each independently sampled from the population of vortices
 removed during the first ten years of UNSEED integration. When a sampled latitude falls in the opposite hemisphere of the
 current warm season, the latitude sign is flipped to follow the seasonal hemispheric restriction while preserving the latitudinal
 distribution of interventions. To maintain symmetry with the moisture decrements applied during vortex removal in UNSEED,
 270 moisture is injected into the q fields in SEED using the same $2.5\times$ scaling factor.

All model experiments were run for 15 years, except SEED_EX, which was integrated for 10 years. Experiments were
 branched from identical initial conditions on 1 February of Year 0005.

2.5 TC footprinting

To quantify the direct contribution or “footprint” of TCs to the modeled precipitation, surface heat and momentum exchange,
 275 and advective flux fields, a TC footprinting technique is devised by masking grid cells associated with TCs in 6-hourly in-
 stantaneous model output. This masking is carried out by the `NodeFileFilter` and `NodeFileEditor` routines of Tem-



pestExtremes. In the literature, the TC outer radius is often defined using a threshold for the azimuthal-mean tangential wind speed (e.g., a surface wind speed threshold of 8 m s^{-1} defines r_8 in Schenkel et al., 2017) or a fixed radial distance (e.g., 1000 km in Arakane and Hsu, 2021). The tangential circulations of TCs largely dictate their direct contribution to horizontal transports and surface aerodynamic fluxes. After testing various wind speed thresholds at the lowest model level ($z \approx 60 \text{ m}$) for correspondence to the region of TC tangential circulations and rainbands, we selected 4 m s^{-1} . We denote this radius r_4 (R4) and use it to compare local circulation anomalies with large-scale responses in Sect. 3.3.

3 Results

3.1 Response of TC climatology to manual TC modulation

To visualize the TC climatology of AQUA, Fig. 4 shows the latitudinal migration of the SST maximum (green dots) and TC genesis density (shading) throughout the year. Genesis is defined as the earliest 6-hourly TC fix along each TC trajectory. In CTL (Fig. 4b), the TC genesis bands migrate with the latitude of maximum SST, remaining just poleward of it. In the Northern Hemisphere (NH), a latitudinally wide seasonal maximum (purple and orange shading spanning $10\text{--}20^\circ\text{N}$) in TC genesis occurs between July and October, roughly matching the real-world seasonality (Schreck et al., 2014). A second, narrower NH seasonal maximum in TC genesis occurs between December and February, likely owing to the strong double ITCZ bias in AQUA. Both the summertime and wintertime genesis maxima are also seen in the Southern Hemisphere (SH), shifted by six months from their Northern Hemisphere counterparts. The remaining panels will be discussed once the mean CTL climatology has been established.

To leverage AQUA as a simplified model of the observed TC climatology, the remainder of the text focuses on TC activity associated with summer and fall genesis events. There are several important definitions. As introduced in Sect. 2.4, “warm season” is defined as June–November (JJASON) in the NH and December–May (DJFMAM) in the SH. Because the annual-mean forcing of the AQUA configuration is hemispherically symmetric, the two warm seasons can be combined into a single climatology to double the statistical sample. For any seasonally and zonally averaged variable $\psi(\varphi)$, the folded field $\psi_{\text{folded}}(\varphi)$ is defined as:

$$\psi_{\text{folded}}(\varphi) = \frac{\psi_{\text{JJASON}}(\varphi) + \psi_{\text{DJFMAM}}(-\varphi)}{2}. \quad (1)$$

The latitude inversion ($-\varphi$) folds SH summer onto the NH and NH winter onto the SH; for hemispherically antisymmetric quantities (e.g., meridional velocity), the sign of the field is also flipped when reversing the latitudes. Unless otherwise noted, figures labeled “JJASON” show $\psi_{\text{folded}}(\varphi)$, where NH latitudes ($\varphi > 0^\circ$) represent the warm-season composite constructed from NH JJASON and SH DJFMAM, and SH latitudes ($\varphi < 0^\circ$) represent the cool-season counterpart.

Figure 5 shows annual counts of TCs and hurricanes as well as accumulated cyclone energy (ACE) for all experiments. The central column contains the annual sum of warm-season TC counts for CTL. Compared to the real world, the warm seasons alone in CTL produce more discrete TCs and hurricane-strength TCs as well as ACE than annually observed (Schreck et al., 2014). Still, with TC number and ACE on the same order of magnitude as observed, CTL serves as a reasonable baseline

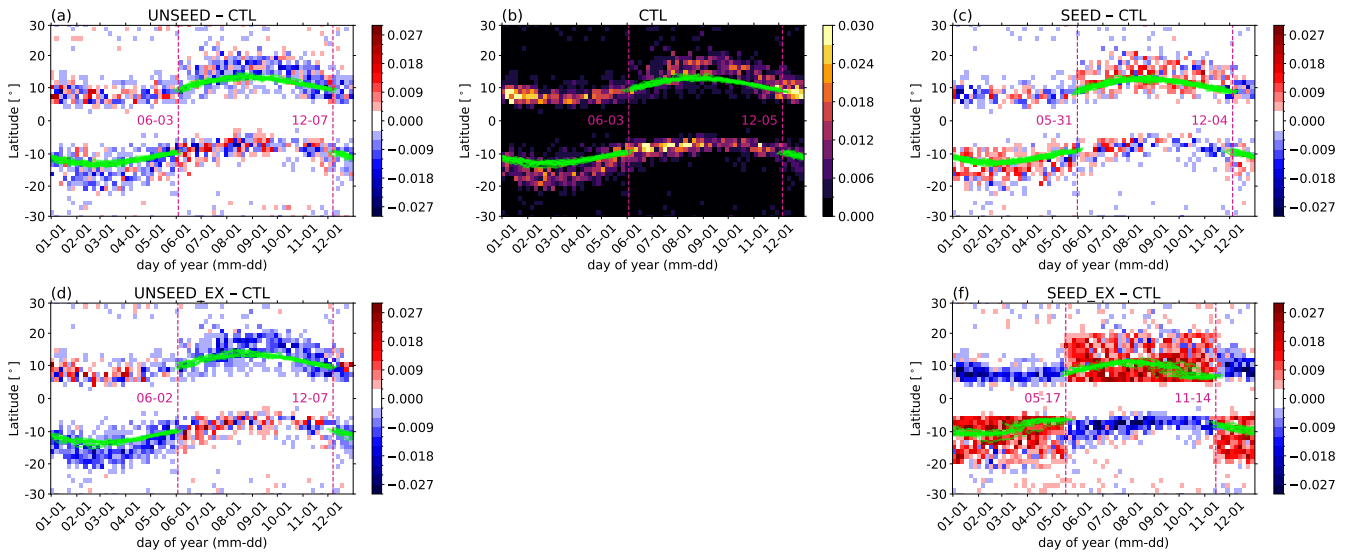


Figure 4. TC genesis events binned by day of year (x -axis) and genesis latitude (y -axis). Shaded values are normalized to density units of $[\text{storms} \times (\text{yr} \times (\text{day of year}) \times 10^6 \text{ km}^2)^{-1}]$. Panel (b) shows the density for CTL, and differences from CTL are shaded in all other panels. The panels are arranged in an upside-down ‘U’ shape, following the ranking from lowest to highest TC frequency clockwise: (d), (a), (b), (c), (f). Green dots show the latitude of maximum zonal-mean SST for each calendar day over the 15-year simulations (10 years for SEED_EX). In magenta, the dashed vertical lines and dates printed in each panel show the mean day of year when this maximum switches hemispheres.

for global TC activity. Fig. 6 shows the lifetime maximum intensity of TCs in each experiment as measured by the minimum central SLP. As before, restricting attention to Fig. 6b for now, most storms are weak in CTL, with a minimum SLP greater than 1000 hPa. The number of storms tapers gradually with increasing intensity, and only a handful of storms intensify toward the lowest theoretical pressures (panel e) predicted by potential intensity (PI) theory (Emanuel et al., 2004; Bister and Emanuel, 2002).

To visualize the life cycles and movement of TCs in aggregate, the warm-season-mean latitudinal distribution of TC activity is displayed in the NH of Fig. 7. Again, focusing on CTL (panels e-f), the warm-season TC genesis frequency (orange line) peaks around 15°N and the TC lysis frequency (blue line) peaks at 20°N , indicating a net poleward migration of storms. The peak in total TC track density (black line in Fig. 7f showing 6-hourly TC fixes) lies between the genesis and lysis peaks. The region of greatest ACE is located just poleward of the lysis peak, indicating storms at those latitudes are disproportionately intense. The cool-season development region in CTL (i.e., negative latitudes) produces a large but narrow genesis peak (high orange line at negative latitudes), which results in a comparable number of TCs (131.8 per year) to the warm season. These storms undergo lysis at lower latitudes (reduced spread between orange peak and blue peak in Fig. 7e) and are generally weaker (lower ACE in Fig. 7f).

Having established the warm-season TC climatology in CTL as an analog to the real world, we next examine how seeding and unseeding alter TC activity. Moving back to Fig. 4, the frequency and spatial density of tracked TC genesis events are

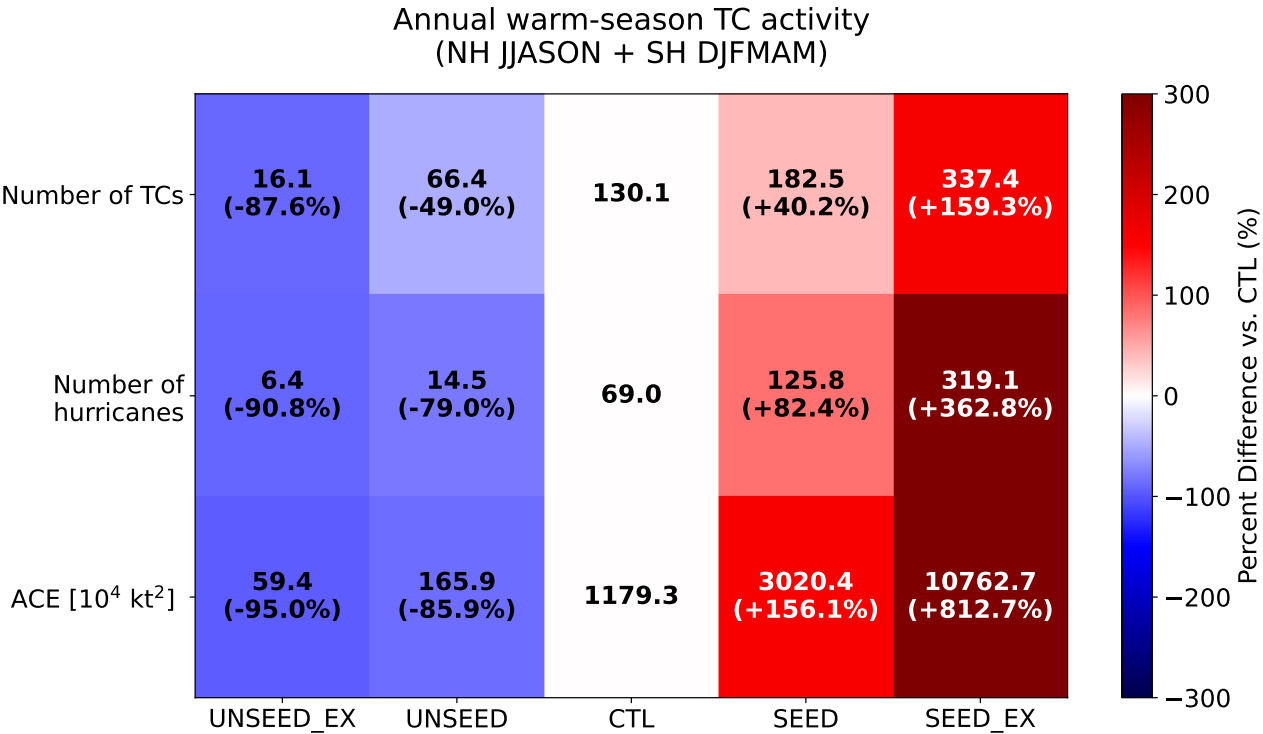


Figure 5. Annual counts of TCs and hurricanes as well as total ACE for warm-season TCs (sum of Northern Hemisphere June–November and Southern Hemisphere December–May). TCs are binned into seasons according to their month of genesis. Absolute counts and ACE are printed in each cell. The percent change relative to CTL is listed in parentheses and shaded.

shown as anomalies relative to CTL, such that adding the anomalies in the side panels to panel b recovers the total genesis frequency. The unseeding experiments (left panels) yield a reduction in TC genesis frequency in the warm season (e.g., the predominantly blue shading in the Northern Hemisphere during JJASON, bounded by the vertical magenta dashed lines. SEED (panel c) produces a TC genesis anomaly with roughly the opposite sign (red shading in the warm season) and comparable magnitude to UNSEED (panel a), consistent with the intended symmetry of the experimental design. SEED_EX produces a substantially larger response, with warm-season genesis enhancement concentrated within the 5–20° latitude seeding band. The more extreme runs (_EX) also exhibit an opposite-signed cool-season response, though this is most apparent in SEED_EX.

These changes in the seasonal genesis distributions are reflected in Fig. 5: unseeding reduces climatological TC frequency and ACE (blue shading), whereas seeding increases both (red). Relative to CTL, UNSEED reduces the total TC count by approximately one-half. The comparatively larger decreases in hurricane count and ACE in the bottom two rows show that the reduction in TC counts is concentrated among the strongest storms, consistent with the shift to weaker lifetime maximum intensities (smaller left tail) in Fig. 6a. Using less restrictive online detection thresholds, UNSEED_EX achieves a further reduction in TC count but only modest additional decreases in hurricane count and ACE. The additional interventions in UNSEED_EX

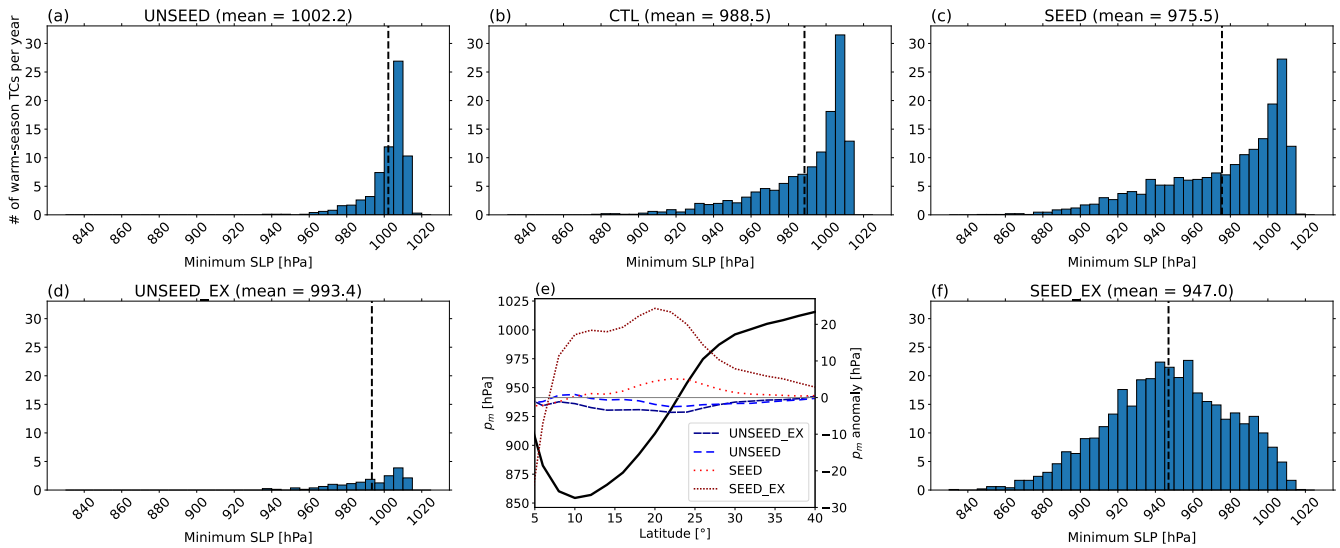


Figure 6. Panel (e) shows the “JJASON” (ψ_{folded} ; Eq. 1) averaged tropical cyclone minimum surface pressure p_m predicted by potential intensity theory. The black solid line shows the CTL value by latitude. The blue dashed and red dotted lines show the anomalies for the unseeding and seeding experiments, respectively. All other panels, arranged in the same upside-down ‘U’ shape as Fig. 4, show histograms of lifetime maximum intensity measured as minimum SLP for warm-season TCs (NH JJASON and SH DJFMAM). The vertical axis shows the annual number of storms in each intensity bin. The mean value is printed above each histogram and marked with a dashed vertical line.

therefore appear to act primarily on the residual population of weak storms that remains in UNSEED. The diminishing response is expected because vortex removal can only act on disturbances generated by the model itself. Consequently, the theoretical maximum reduction is the complete suppression of the model’s intrinsic (CTL) TC activity, with progressively smaller gains as this limit is approached.

In contrast, the total effects of seeding are not bounded by the baseline TC activity. With an imposed forcing designed to statistically oppose UNSEED, SEED produces nearly equal and opposite changes in TC frequency and hurricane frequency in Fig. 5. However, a stark asymmetry emerges in the integrated intensity: ACE increases by more than 150% relative to CTL, exceeding the theoretical maximum decrease attainable through unseeding. To probe this response, SEED_EX is run with more frequent interventions (1 per day) and larger initial vortex amplitudes (Table 1). This extreme case produces more than double the TC count of CTL, with increases in ACE and hurricane frequency several times larger than those of all other experiments. The disproportionate enhancement of ACE in SEED_EX is driven in part by a shift towards higher storm intensities (leftward shift of the distribution in Fig. 6f) and suggests that the tropics in AQUA possess a reservoir of potential energy that spontaneously generated TCs in CTL do not tap into.

Though aggregated TC activity metrics confirm the efficacy of seeding and unseeding, they do not reveal how individual interventions relate to the emergent TC climatology. Comparing intervention locations with offline TC trajectories shows that this relationship is not one-to-one (i.e., one unseeding intervention doesn’t necessarily remove one TC from the tracked record).

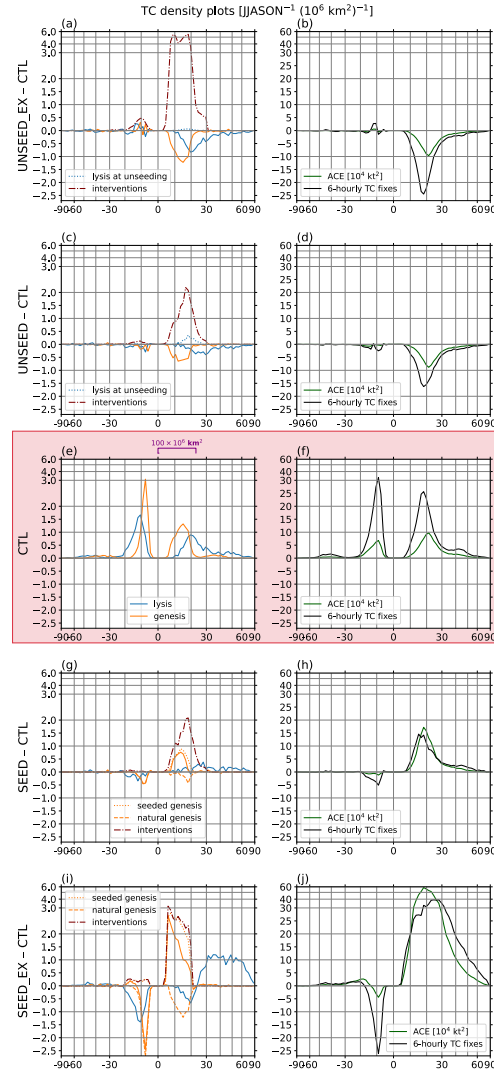


Figure 7. Normalized TC activity and intervention responses for the folded “JJASON” (ψ_{folded} ; Eq. 1) composite, with latitude ($^{\circ}$) on the x -axis. Northern Hemisphere latitudes ($\varphi > 0^{\circ}$) represent the general warm season, while Southern Hemisphere latitudes ($\varphi < 0^{\circ}$) represent the general cool season. Experiments are ordered from top to bottom by increasing warm-season TC frequency. All data are shown as differences from CTL in units of counts (or ACE) per 10^6 km^2 per JJASON-equivalent season (such that integrating a curve across both hemispheres gives the global climatology over a representative JJASON-equivalent six-month season). **Left Panels:** Latitudinal densities of TC genesis (orange) and lysis (blue) points. Maroon lines indicate seeding/unseeding interventions. Blue dotted lines denote lysis points coincident with unseeding interventions. Orange dotted lines indicate genesis points matched to placed seeds (see text for matching criteria), while the orange dashed lines show the net change in natural (non-seeded) genesis relative to CTL. **Right Panels:** Latitudinal density of Accumulated Cyclone Energy (ACE; green) and total 6-hourly TC fixes (black). **Visual Formatting:** The horizontal axis is sin-scaled with gridlines placed every 10° latitude. The vertical axes use a symmetric logarithmic scale with horizontal gridlines placed every 1 unit in the left column and every 10 units in the right column.



In UNSEED, the latitude of unseeding interventions (maroon dash-dotted line in Fig. 7c) broadly coincides with the greatest TC track density in CTL (black line in Fig 7f). Compared to CTL, UNSEED exhibits reduced TC genesis (orange line), lysis (blue solid line), ACE (green line), and TC fixes (black line). The climatological distributions alone invite multiple interpretations. One possibility is that unseeding suppresses TCs throughout their life cycle. Another is that it primarily prevents incipient disturbances from reaching TC genesis criteria, with the downstream reductions in ACE, lysis, and track density necessarily following.

Distinguishing between these possibilities requires understanding the relationship between the online intervention algorithm and the offline tracking methodology (Sect. 2.2). The online unseeding algorithm operates every 24 hours on any snapshot feature that meets the local `DetectNodes` criteria, effectively acting on the total population of tropical disturbances. However, in our offline climatology, `TempestExtremes` filters out these candidate features unless they successfully trace a trajectory spanning at least 10° GCD and sustain wind speeds of at least 10 m s^{-1} across ten 6-hourly steps. Thus, if unseeding preferentially targets disturbances shortly after their emergence, these features are likely to be removed before they satisfy the persistence and intensity thresholds required for offline TC tracking. This reasoning predicts that relatively few interventions should coincide with the lysis of offline-tracked TCs.

This is borne out by comparing interventions with TC trajectories. Only a small fraction of interventions coincide with the lysis point of an offline-tracked TC (blue dotted line in Fig. 7a,c), indicating that direct TC termination is not the dominant outcome of unseeding. Furthermore, the partitioning of unseeding interventions in Fig. 8c reveals that nearly 80% of all interventions in UNSEED target the transient candidate vortices (large yellow-green pie-chart sector) that never match a TC fix. These interventions may therefore either represent successful suppression of disturbances that would otherwise have developed into TCs or reflect the targeting of short-lived features that would not have satisfied the offline TC criteria even without intervention. Only the remaining one-fifth of interventions target TCs that achieved the thresholds required to appear in the offline record. Together, these findings support the interpretation that unseeding predominantly prunes incipient disturbances before they can be tracked as TCs, rather than primarily terminating mature storms.

This partitioning is even more extreme in UNSEED_EX: over 98% of interventions target candidate vortices that do not become offline-tracked TCs. Although the intervention rate (647.8 per year) is more than triple that of UNSEED (212.3 per year), a smaller absolute number of them are collocated with a lysis point (near-zero blue dotted line in Fig. 7a). Thus, increasingly aggressive unseeding (achieved through broader online detection criteria) shifts the algorithm towards suppressing disturbances before they meet the trajectory length and duration thresholds of `TempestExtremes`, rather than terminating additional tracked TCs. Compared to UNSEED, the additional 435 interventions per year in UNSEED_EX only yield a further reduction of approximately 50 TCs per year, suggesting that many additional interventions act on candidate vortices with limited contribution to the diagnosed TC climatology. UNSEED_EX therefore demonstrates a practical limit of this unseeding algorithm: beyond a certain intervention frequency, additional vortex removals produce diminishing reductions in diagnosed TC activity.

Despite the large number of unseeding interventions in UNSEED_EX and UNSEED, a portion of warm-season TCs still develop. Nearly half of the tracked warm-season TCs in UNSEED and over 90% in UNSEED_EX encounter zero unseeding interventions (large blue pie-chart sectors in Fig. 8b,d). This structural omission primarily stems from the differing online

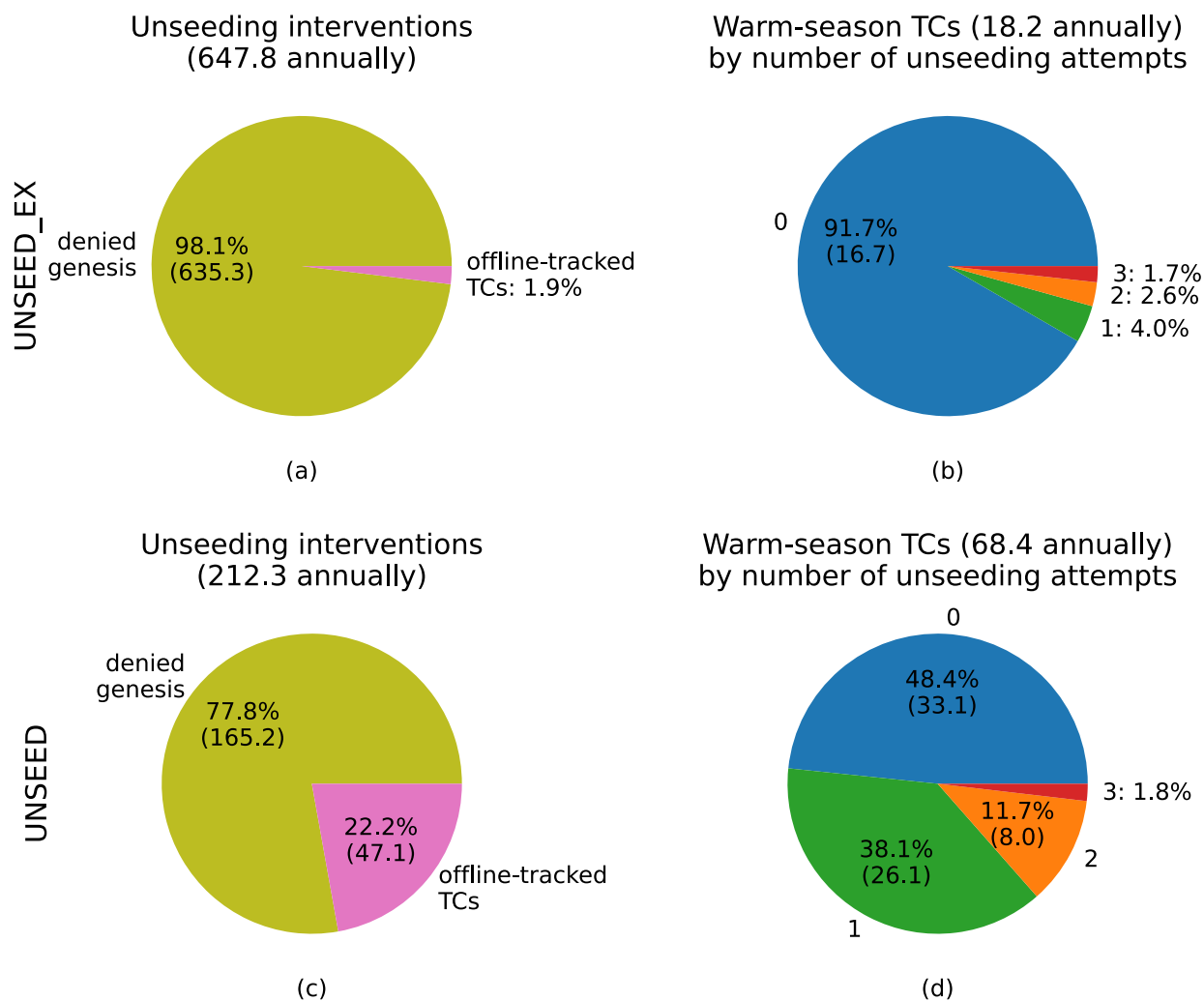


Figure 8. Pie charts depicting the allocation of unseeding interventions (left) and the count of warm-season TCs by the number of unseeding attempts along their trajectories (right). The right-hand panels include any TC with at least one 00Z TC fix in the same hemisphere in which unseeding is active (i.e., the hemisphere of maximum of zonal-mean SST). Percentages are printed within or adjacent to each slice, with the corresponding annual counts shown in parentheses where space permits. The large olive-green slices labeled “denied genesis” in the left panels represent unseeding interventions that do not match any TCs detected by TempestExtremes offline.



and offline detection criteria detailed in Appendix A. Specifically, restricting the online W_{OFFSET} threshold to 0.5° GCD
 390 (compared to 1° GCD offline) ensures that the upright, axisymmetric inverted RJV perturbation is only applied to vertically
 aligned vortices. Consequently, nascent disturbances characterized by vertical tilt or environmental shear slip through the online
 filtering framework, yet can be captured by the more permissive offline detection.

For the TCs that *are* targeted, the unseeding framework is effective. In UNSEED, most unseeded storms undergo rapid lysis
 after a single intervention (green sector), confirming that the inverted RJV and localized moisture removal successfully suppress
 395 the candidate vortex. A smaller subset (9.2 TCs per year in the orange and red sectors) requires multiple interventions before
 undergoing lysis. This recalcitrance suggests that the axisymmetric analytical perturbation does not fully neutralize storms
 with asymmetric structure or atypical radial profiles. This trend is amplified in UNSEED_EX, where a larger fraction of storms
 experience multiple unseeding events relative to one unseeding event. Because UNSEED_EX targets weaker disturbances,
 greater geometric mismatch between the actual TC and the fitted RJV profile could explain why some trajectories require 2–3
 400 interventions for termination. Collectively, these statistics demonstrate that while structural thresholds allow a portion of TCs
 to initially bypass real-time intervention, the unseeding framework remains fairly efficient at terminating a TC trajectory once
 an intervention is triggered.

Unlike unseeding, seeding interventions are straightforward to link to TC life cycles. A successful seeding intervention
 is defined as one that matches a nearby (within 3° GCD) genesis point occurring exactly one or two 6-hour model output
 405 intervals later, as the seeded vortex converts into a trackable TC. By design, the latitudinal distribution of seeds in SEED nearly
 matches the pattern of unseeding interventions in UNSEED (maroon lines in Fig. 7c,g). From Table 2, the seed conversion
 rate of 33.4% means that only one-third of seeds match a genesis point in the offline record. Most of these successful seed
 conversions (orange dotted line in Fig. 7g) correspond to seeds placed in the equatorward half of the seeding band. Fig. 6e
 shows this is where the background state offers inherently greater environmental support as estimated by PI minimum pressure
 410 p_m (solid black line). The response efficiency is fairly high: the 71.4 annual successfully seeded TCs lead to a net increase of
 52.4 discrete TCs per year. Under moderate seeding, seeded TCs approximately stack onto the background TC climatology
 with limited mutual suppression.

In contrast to the low conversion rate in SEED, the high-amplitude and radially large seeds in SEED_EX achieve a near-90%
 conversion rate, with almost all seeds surviving across the seeding region (the orange dotted line lies very close to the maroon
 415 line in Fig. 7i). The strong seeded TCs come at the expense of most naturally generated TCs (large negative values of the
 orange dashed line), resulting in a lower response efficiency of 41.8%. This hints at a reduction in environmental support for
 TCs compared to CTL, consistent with the PI (higher p_m shown by the positive dark-red dotted line in Fig. 6e). In addition
 to the warm-season responses, the cool-season TC count is drastically reduced in SEED_EX, implying substantial modulation
 via cross-equatorial energy transports or a memory effect (e.g., Hart et al., 2007). The contrast between SEED and SEED_EX
 420 demonstrates that algorithmic tuning parameters do not translate directly into TC activity changes: the ambient environment
 regulates seed survival, while the resulting storms modify the environment for subsequent TCs.



	SEED	SEED_EX
Seed conversion rate	33.4%(71.4 / 213.9)	89.2%(325.1 / 364.6)
Response efficiency	73.4% (52.4 / 71.4)	41.8%(135.8 / 325.1)

Table 2. Average annual metrics for the SEED and SEED_EX experiments. (Top row) *Seed conversion rate* = $N_{\text{prox}}/N_{\text{seed}}$, where N_{prox} is the number of TC genesis points sufficiently close to a seed in space/time (see text), and N_{seed} is the total number of seeding interventions. (Bottom row) *Response efficiency* = $\Delta\text{TC}/N_{\text{prox}}$, where ΔTC is the net increase in the total number of discrete TCs.

3.2 Response of the tropical mean state

We now turn to the mean climate response. Examining mean fields allows us to evaluate AQUA as a tropical testbed for isolating interactions among TCs, energy budgets, and the large-scale circulation. It is important to note that because warm-season interventions are applied continuously and frequently in our experiments, the tropical atmosphere (even with its short relaxation timescales) does not fully relax back to its unperturbed baseline. The resulting time-mean anomalies instead represent TC-maintained climate states evolving toward a perturbed statistical equilibrium and may therefore qualitatively differ from composite studies of environmental recovery after individual TCs. Our objective here is to document the robust large-scale responses enabled by the intervention framework, rather than to fully attribute the processes producing each response.

Figures 9 and 10 list area-averaged, time-mean quantities for the warm-season tropics and subtropics (10–30° latitude), with a vertical mass-weighted integral $\int(\cdot)\frac{dp}{g}$ applied to energy content and cloud-water metrics up to 100 hPa (nominally through the depth of the troposphere). The poleward limit of 30°N was chosen based on the canonical outer edge of the tropics, and the equatorward limit of 10°N excludes potential contamination by AQUA’s equatorial cold tongue. The 10–30° latitude band in the warm season also aligns with the region of greatest intrinsic ACE and perturbations to ACE in our experiments. To avoid interpolation errors, both the regional area-weighted means presented here and the following zonally averaged fields are calculated directly on the native, unstructured ne120np4 cubed-sphere mesh using the UxArray library (Chen et al., 2026).

The tabulated metrics yield a zeroth-order picture of how persistent TC activity anomalies manifest across basin scales. ACE represents the time-integrated intensity of tracked storms, whereas EKE measures the variance of the resolved atmospheric flow capturing transient TC circulations and other disturbances. Across all five simulations, a strict monotonic relation emerges between TC activity and EKE (i.e., more ACE correlates with greater EKE from left to right). This relationship is consistent with TC activity contributing to the resolved transient circulation field, although monthly mean EKE cannot isolate that direct contribution from other TC-induced dynamical adjustments.

As TC activity increases from left to right in Fig. 9, the tropical-mean SST decreases monotonically (first row). This correlation aligns with previous studies showing local SST cooling due to TC wind forcing (Sriner and Huber, 2007; Li et al., 2023) and basin-scale upper-ocean cooling with greater TC activity (Hart et al., 2007; Schenkel and Hart, 2015b; Rios et al., 2026).

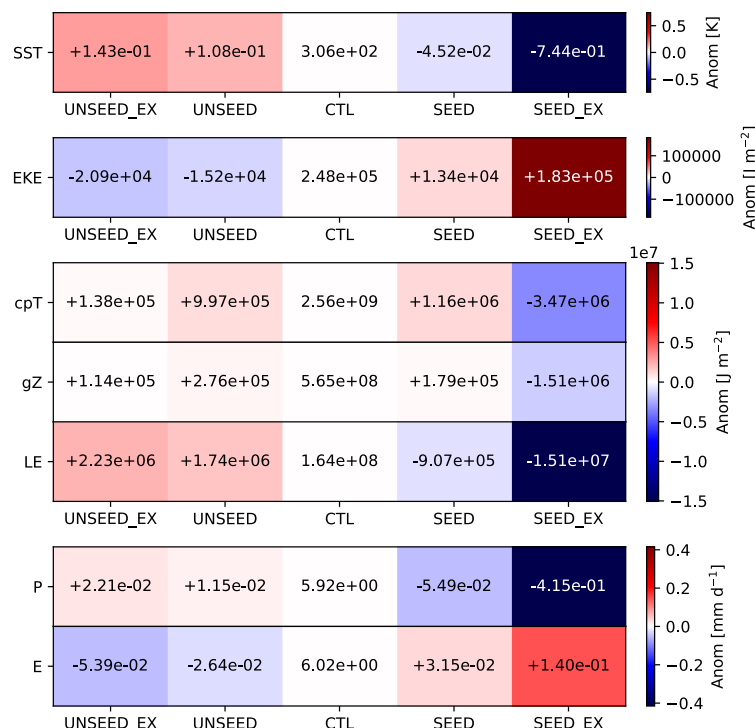


Figure 9. Area- and time-averaged tropical mean state metrics for 10–30°N JJASON and 10–30°S DJFMAM (i.e., the warm season). TC activity increases from left to right. Dry static energy (DSE) is the sum of sensible heat $c_p T$ and geopotential gZ . Moist static energy (MSE) equals the sum of DSE and latent energy (LE). The atmospheric energy content metrics (EKE, DSE, and LE) are vertically integrated from the surface up to 100 hPa. CTL shows the raw averages. For unseeding and seeding experiments, anomalies relative to CTL are indicated by both cell shading and printed values. Contiguous rows with no vertical spacing share a common colorbar.

Similar to SST, the tropical-mean latent energy (LE) associated with water vapor decreases as TC activity increases relative to CTL, even though we artificially remove water vapor in both unseeding runs and add water vapor in SEED. This TC-associated drying matches basin-scale behavior in the composite-based study of Schenkel and Hart (2015a).

Just as the LE shows a clear relationship to TC activity, so do surface hydrological fluxes of precipitation P and evaporation E . With greater TC activity in AQUA comes a reduction in 10–30°N precipitation. Given that TCs themselves produce high local precipitation rates, the mean response indicates an overcompensation by the background non-TC state: mean precipitation is reduced by some combination of tropospheric drying and weaker total background convective fluxes in response to TCs. This suppression of rainfall is clear evidence that the total climate response is not a global accumulation of anomalous conditions within TCs. Instead, TCs affect their environment beyond the outer radii and lifespans of individual storms, and our experimental framework captures the long-term, tropical-mean expression of these upscale impacts. On the other side of the hydrological budget, mean evaporation increases with TCs. The simple explanation is that strong surface winds in TCs induce

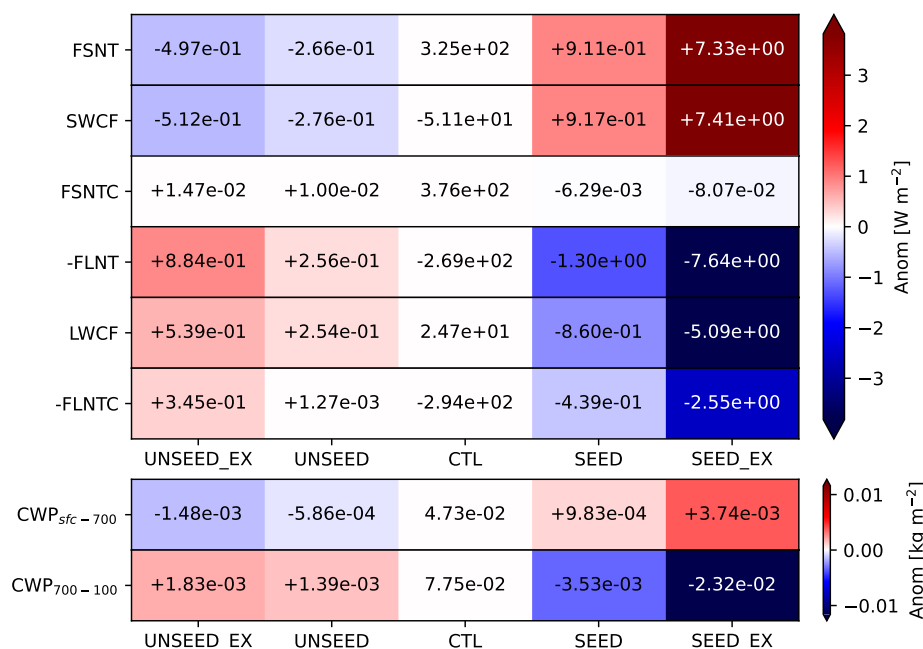


Figure 10. Area- and time-averaged TOA radiative fluxes and cloud water paths for 10–30°N JJASON and 10–30°S DJFMAM (i.e., the warm season). FSNT and FLNT denote net TOA shortwave and longwave fluxes, where the suffix C indicates clear-sky fluxes. SWCF and LWCF denote shortwave and longwave cloud radiative forcing. Downward (earth-warming) values are plotted as positive for radiative fluxes and cloud forcing. CTL shows raw averages, while other columns show anomalies. Contiguous rows with no vertical spacing share a common colorbar.

anomalously large evaporative fluxes, but analysis of our experiments indicates that anomalies in mean trade winds and SSTs also play a large role in dictating E (not shown), which merits further study in the future.

In contrast to the consistent sign changes of the EKE and LE with TC activity, the tropical-mean tropospheric dry static energy (DSE) does not display a clear trend or sign of anomalies across the experiments. Whereas SEED_EX and UNSEED indicate a marked inverse relation of DSE to TC activity, SEED shows the opposite, though changes in UNSEED_EX are small. These minor inconsistencies may arise from internal variability or feedbacks involving radiative fluxes and the mean circulation.

The lack of a tropical-mean DSE signal warrants examination of spatial patterns and thermodynamic budgets. The atmospheric MSE (shading) and DSE (contours) are plotted in Fig. 11 atop the upper-ocean heat content (shaded as potential temperature θ_{ocn}). The basic state in CTL (panel b) features a stably stratified atmosphere (increasing DSE with height), high moisture content at the double ITCZs (latitudinal MSE maxima below 400 hPa just poleward of 10°N and equatorward of 10°S), and a well-defined warm-season thermocline (strong vertical gradient of potential temperature at ~50–100 m depth) separating warm tropical surface waters from cooler waters beneath.

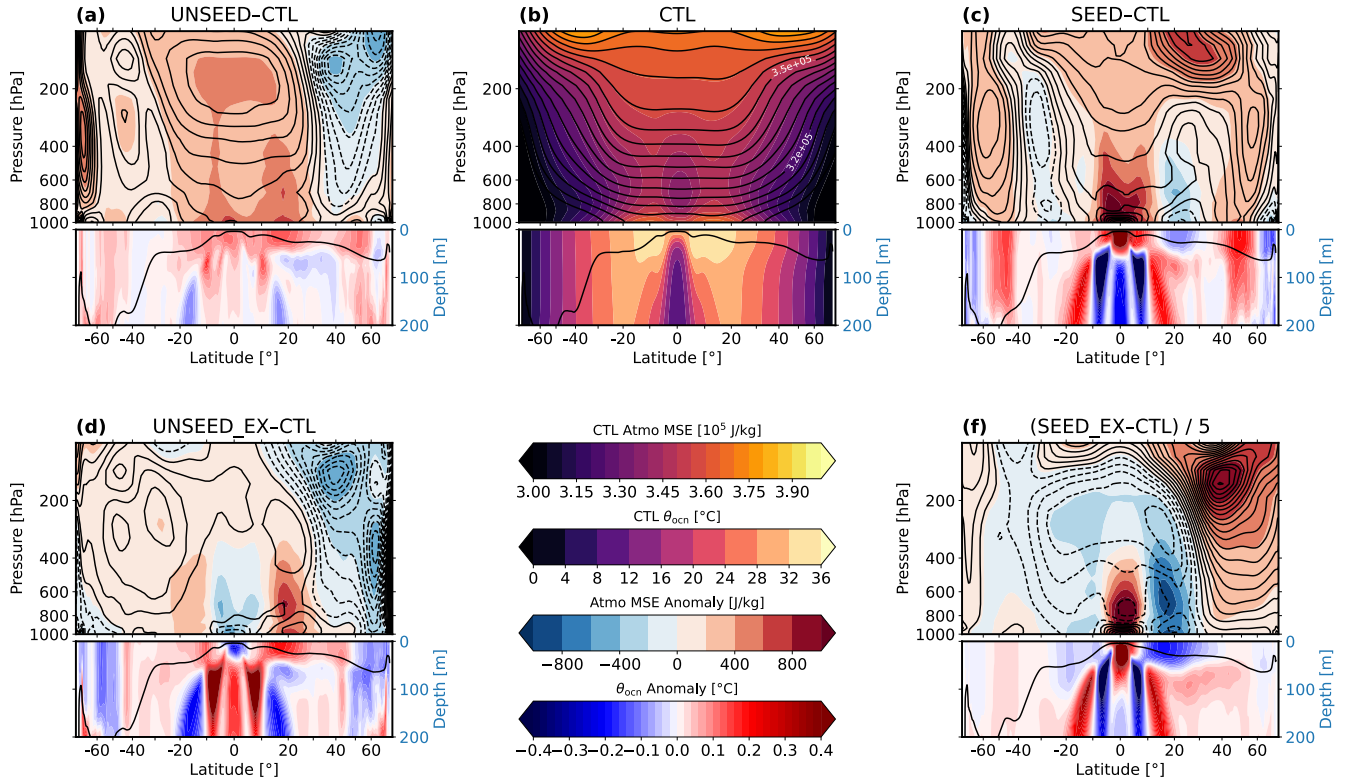


Figure 11. Zonally averaged JJASON (ψ_{folded} ; Eq. 1) thermodynamic state of the troposphere and upper ocean, where positive latitudes correspond to the warm-season mean and negative latitudes correspond to the cool-season mean. Within each panel, the vertical axis is split into atmospheric pressure (upper) and ocean depth (lower). In the atmosphere, dry static energy (DSE) is contoured, and moist static energy (MSE) is shaded. In the ocean, potential temperature θ_{ocn} is shaded, and the absolute ocean mixed layer depth (MLD) is denoted by a black curve. DSE contour intervals are 5000 J kg^{-1} in (b) and 50 J kg^{-1} in all other panels. DSE, MSE, and θ_{ocn} show differences from CTL.

470 Focusing first on the ocean state, all perturbed runs produce a vertical dipole in ocean potential temperature anomalies roughly spanning $10\text{--}30^\circ\text{N}$ between the surface and 100 m depth. In SEED and SEED_EX (Fig 11c,f), cold anomalies (blue above ~ 50 m) at the ocean surface overlie warm anomalies (red between 50 and 100 m). This is consistent with the cumulative imprint of enhanced TC-driven turbulent mixing and entrainment. Compared to the seeding runs, UNSEED and UNSEED_EX (Fig. 11a,d) produce the opposite-signed dipole (warm over cold), implying reduced upper-ocean mixing when TC activity is

475 artificially suppressed.

The upper-ocean thermal anomalies are directly communicated to the atmospheric boundary layer through surface turbulent heat fluxes. In all perturbed experiments, the near-surface MSE anomalies share a sign (same shaded color) with the underlying θ_{ocn} at nearly all latitudes. The free tropospheric state (above approximately 850 hPa) is less tightly coupled to SSTs than the boundary layer. Focusing on SEED_EX (Fig. 11f), the tropical free troposphere cools (dashed contours show negative DSE anomalies) equatorward of 30° in both hemispheres, with relative horizontal uniformity resulting from weak-temperature-

480



gradient dynamics. The cool anomaly maximizes in amplitude around 200–300 hPa. In the Northern extratropics (poleward of 30°N), a troposphere-deep warm anomaly exists in SEED_EX, tilting equatorward with height over the subtropics. As expected, UNSEED and UNSEED_EX (Fig. 11a,d) mirror this structure with the opposite sign (tropical warming and boreal extratropical cooling). Overall, the DSE responses in Fig. 11 tend to imply cooling of the tropical troposphere and stabilization
485 of the upper-troposphere/lower-stratosphere region with greater TC activity.

In many regions, the lined DSE and shaded MSE anomaly contours are parallel or coincident, indicating that moisture contributes little independently of temperature to the MSE response. The notable exception (apart from the equatorial cold-tongue bias region) is the TC belt between 10 and 30°N, a region where LE anomalies due to water vapor are large, affecting MSE but not DSE. In SEED and SEED_EX, negative MSE anomalies far exceed negative DSE anomalies over the 400–800 hPa
490 layer between 10°N and 30°N, implying a large reduction in water vapor in this region. UNSEED and UNSEED_EX mirror this MSE anomaly, with mid-tropospheric moistening as TC activity decreases.

These tropospheric and upper-ocean thermal anomalies directly regulate the thermodynamic limit of TC intensity according to potential intensity (PI) theory. By combining the shifts in SSTs with the upper-tropospheric temperature anomalies, one can explain the potential intensity (PI) responses introduced earlier in Fig. 6e. PI is computed as in Bister and Emanuel (2002),
495 with p_m derived from monthly mean model output following Eqs. 3–4 via Gilford (2021). Inputs include column SST, surface pressure, and temperature and specific humidity profiles to 50 hPa, assuming $C_k/C_D = 0.9$. In the seeding experiments, the natural TC genesis frequency decreases (negative orange dashed lines in Fig. 7g,i). This implies that seeded TCs are making the large-scale environment less favorable for other TCs to spontaneously form. At those same latitudes, PI theory shows reduced thermodynamic support for TCs as higher (weaker) p_m (positive red dotted lines in Fig. 6e). This thermodynamic hostility has
500 contributions from SSTs and tropospheric temperature and moisture profiles, including upper-troposphere/lower-stratosphere stabilization. Conversely, the unseeding experiments have stronger PI (negative p_m anomalies shown by blue dashed lines in Fig. 6e), indicating greater thermodynamic support for TCs. Because unseeding removes TC-like disturbances as they form, the resulting TC climatology cannot reveal whether the more favorable environment would otherwise support greater storm frequency. Confirming this feedback therefore requires tracking criteria that capture weak or incipient disturbances before they
505 are removed.

Because mid-tropospheric water vapor strongly affects tropical clear-sky longwave radiation, we next examine the domain-mean top-of-atmosphere radiative anomalies (Fig. 10). Across the five experiments, -FLNT decreases monotonically with increasing TC activity (reddest to bluest shading from left to right). Because the net longwave radiative flux FLNT is defined as positive upward, smaller values of -FLNT correspond to greater net longwave radiation leaving the atmosphere. This net
510 longwave cooling with TCs is driven mostly by reduced cloud fraction and/or lower cloud tops, as shown by a same-signed but smaller magnitude response in longwave cloud forcing (LWCF). The remainder of the longwave cooling comes from the clear-sky component (negative values of -FLNTC). As Rios et al. (2026) noted, the mean SST cooling induced by TCs should weaken clear-sky longwave cooling through the Planck feedback. Because our experiments (and theirs) show a net increase in clear-sky longwave cooling instead, the drying of the mid-troposphere must dominate the clear-sky longwave response. On
515 the other hand, the net shortwave radiative flux at the top of the model (FSNT) contributes to tropical warming as TC activity



increases. This shortwave warming is driven by shortwave cloud forcing (SWCF), with the clearsky component (FSNTC) offsetting the response by $\sim 1\%$. Though FSNT and -FLNT anomalies largely cancel, the net longwave terms prevail in all perturbed experiments except UNSEED, demonstrating a net top-of-atmosphere cooling response to enhanced TC activity in the tropics.

520 The cloud water path (CWP) is tightly associated with the mean convective behavior while also explaining the cloud radiative forcing anomalies in the preceding paragraph. Warm-season area-averaged CWP values ($10\text{--}30^\circ$ latitude) show a monotonic relationship with TC activity (bottom two rows of Fig. 10). As TC activity increases, the long-term tropical-mean CWP of low clouds (surface to 700 hPa) increases, whereas the CWP of medium and high clouds (700 to 100 hPa) decreases. The loss of free-tropospheric cloud water is consistent with reduced deep-convective coverage or intensity, suggesting that greater TC
 525 activity suppresses background convection. Because both LWCF and SWCF weaken, the reduction in mid- and upper-level clouds dominates the net cloud-radiative response.

These mean state and radiative responses demonstrate that AQUA is a useful testbed for isolating coupled climate adjustments to persistent changes in TC activity. Its simplified geography allows these responses to be interpreted as large-scale feedbacks rather than basin-specific variability. Crucially, online vortex modulation alters TC activity while allowing the cou-
 530 pled atmosphere-ocean system to adjust freely.

Radiative fluxes, along with other energy budget responses, serve as a stepping stone to explain the large-scale circulation response of the tropics to TCs (Rios et al., 2026). This raises a central question: does the mean circulation adjust linearly in response to direct energy and momentum transports by TCs, or nonlinearly in response to transports mediated by tropical background convection, which is indirectly controlled by TCs through longer-lived thermodynamic modifications of the
 535 environment?

3.3 TC footprints do not explain the mean circulation response

Working within a linear view of TC upscale impacts on climate, past observational studies (Hsu et al., 2008; Arakane and Hsu, 2021) have shown that thermodynamic and circulation anomalies associated with TCs contribute significantly to the atmospheric mean state and variability at seasonal and longer timescales. To quantify this influence (or so-called TC “footprints”) in reanalysis output, Arakane and Hsu (2021) removed potential vorticity (PV) anomalies within a 1000 km radius from each
 540 TC center and performed PV inversion to adjust other state fields while maintaining balance. From this TC vortex extraction analysis, they concluded that TCs leave detectable imprints in monthly mean atmospheric states commonly interpreted as representations of large-scale, low-frequency climate variability, with implications for diagnostics such as genesis potential indices. A unique aspect of our work is that our TC denial experiments permit us to assess how accurately this vortex-removal
 545 framework describes a “no-TC” climate, thereby testing whether TCs interact linearly or nonlinearly with the mean circulation.

This linearity is tested in Fig. 12 by comparing the change in TC-induced surface stress to the change in mean surface stress. In CTL (panel b), the mean surface stress (blue line) is easterly (negative), peaking between 20°N and 30°N , consistent with the observed zonal mean. The TC-contributed stress (orange line) is computed by first masking each instantaneous zonal surface stress field so that all grid points outside the TC r_4 (or R_4 ; see Sect. 2.5) radius are set to zero, and then taking the same



temporal and zonal mean as for the full field. In other words, this is the direct contribution of TC footprints to the climatological zonal stress, and in CTL it is also net easterly within the region of easterly trade winds.

With more intense and frequent TCs in SEED and SEED_EX (right panels), the TC footprints contribute to increased (anomalous relative to CTL) easterly stress over 20–30°N (Fig. 12c,f; negative orange lines). However, the mean circulation response (blue lines) at the same latitudes shows a positive (westerly) anomaly. This indicates that the mean trade winds weaken with increased TC activity, despite the fact that the diagnosed TC-footprint stress contribution is easterly and therefore opposes the sign of the large-scale circulation response. This opposition provides evidence that the climate response cannot be explained as a basic linear addition of TC momentum forcing to an otherwise fixed background state. The unseeding experiments reinforce this conclusion: reduced TC activity weakens the easterlies induced by TC footprints (Fig. 12a,d; positive orange lines), but the long-term mean easterly stress strengthens (negative blue lines). Just as Hsu et al. (2008) hypothesized, a true “no-TC” climate cannot be approximated simply by masking out storms because the background state has already adjusted to their presence. Instead, the sign reversal shows that the long-term surface-stress response is dominated by adjustment of the large-scale background circulation rather than the direct momentum footprints of TCs.

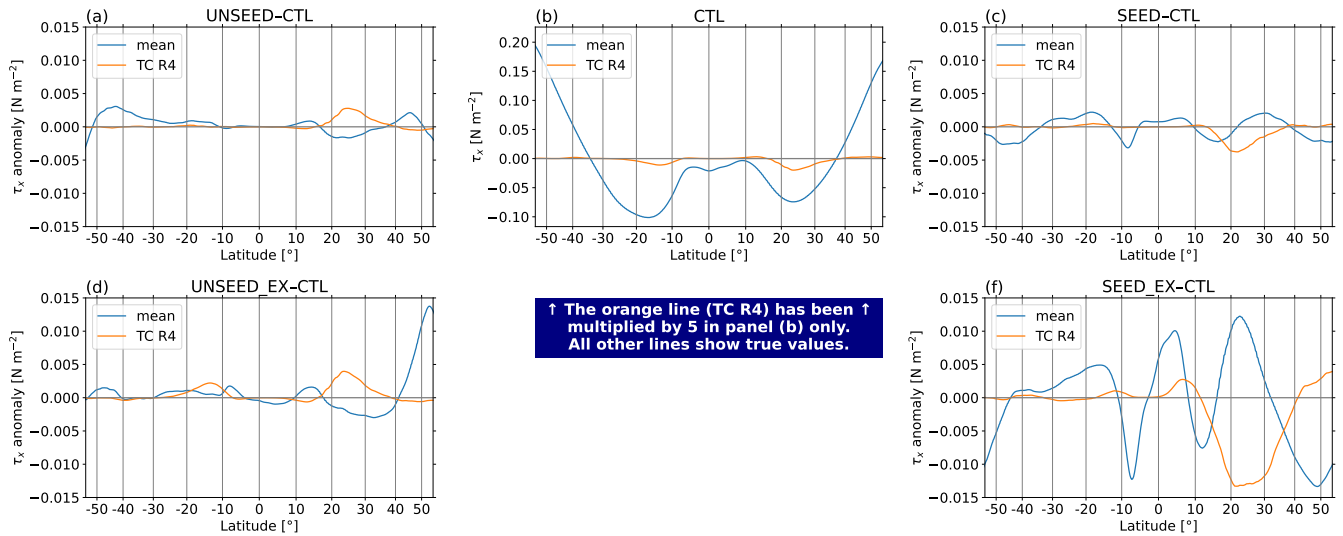


Figure 12. Zonal-mean zonal surface stress τ_x [N m^{-2}] from 6-hourly instantaneous output for years 0006–0007, representing JJASON (ψ_{folded} , Eq. 1). Positive τ_x represents eastward stress experienced by the ocean surface, implying eastward surface winds. Blue lines show the net zonal surface stress. Orange lines show the zonal stress recomputed with stress values outside of TC r_4 set to zero, representing the direct contribution of TC footprints to the total stress. (b) represents the climatological mean of CTL, while the other four panels represent the deviation from this mean for the other experiments. All panels other than (b) share a vertical axis. The orange line in (b) is multiplied by a factor of five to facilitate visual comparison.

The surface-stress response implies an adjustment of the equatorward near-surface branch of the Hadley circulation needed to maintain the zonal momentum balance. Through mass continuity, this adjustment would be accompanied by changes in tropical



565 vertical motion, linking the circulation response to background convective overturning and radiatively driven subsidence. These inferred changes are consistent with the thermodynamic and radiative responses in Sect. 3.2, but their underlying mechanisms require a more detailed budget analysis.

4 Discussion

This study evaluates online vortex insertion and removal as a means of modulating TC activity in a GCM. Unlike offline re-
 570 moval or composite-based approaches, online manipulation allows the coupled atmosphere-ocean system to adjust to altered TC activity and preserves the associated feedbacks. Both seeding and unseeding superimpose axisymmetric RJVs (Reed and Jablonowski, 2011) with tunable structural parameters. Seeding parameters are sampled from prescribed distributions, whereas unseeding parameters are iteratively fitted so that an inverted RJV cancels the targeted candidate vortex. The imposed perturbations retain the local dynamical balance of the Reed formulation, while global field corrections preserve selected atmospheric
 575 integrals and limit artificial budget drift. Across the aquaplanet experiments, seeding increases TC frequency and activity above baseline, whereas unseeding reduces both. The resulting tropical mean-state and radiative responses vary mostly monotonically with TC activity, showing that persistent TC anomalies can alter climate beyond the direct footprints of individual storms. Together, these results establish online vortex modulation as a practical and climatologically stable means of controlling TC climatology, providing a new experimental framework for studying TC upscale impacts in Earth system models.

580 The mechanics of seeding are relatively straightforward to interpret. When a cyclonic RJV is inserted, it either converts into a TC through self-sustaining surface heat flux (Emanuel, 1986) and radiative feedbacks (Wing et al., 2016; Muller and Romps, 2018), or dissipates if the seed amplitude is too small or the environment too hostile. In the case of survival, the seeded vortex typically evolves into an offline-tracked TC, with the resulting genesis point occurring within one or two 6-hour model output intervals and 3° GCD of the seeding intervention. Most seeded TCs stack on top of the baseline TC counts of CTL, indicating
 585 that the tropics of AQUA contain potential energy that internally generated TCs do not tap into. However, the internally generated TCs do decrease, especially as intervention frequency and amplitude are increased in SEED_EX, suggesting that seeded TCs make the environment more hostile for subsequent genesis. This demonstrates that vortex insertion provides strong control over TC frequency while still allowing the ultimate TC climatology to emerge from environmental interactions.

In contrast to the constructive effect of seeding on the TC genesis climatology, unseeding largely acts on the pre-genesis
 590 disturbance population. Only a handful of unseeding interventions terminate mature, offline-tracked TCs. Most instead remove candidate vortices early in their life cycles, before they satisfy the offline TC trajectory criteria. Despite directly terminating only a small number of mature TCs, unseeding is highly effective, eliminating 85–95% of ACE from the unseeding runs. As online detection thresholds are relaxed to increase intervention frequency, as in UNSEED_EX, further reductions in TC activity are achieved by increasingly targeting a large population of weak, incipient disturbances.

595 Residual TC activity during unseeding partly reflects limitations of the current implementation rather than a fundamental limit of vortex removal. First, the 24-hour intervention interval allows disturbances to form and intensify between restarts; central pressures near 970 hPa at some interventions show that substantial development at sub-daily timescales can occur, not



unlike in observations (Kaplan and DeMaria, 2003). Implementing unseeding directly in the atmospheric model would avoid restart overhead and permit more frequent intervention. Second, the online `DetectNodes` criteria are stricter than the offline tracking criteria, particularly for vertically tilted vortices, so some tracked TCs are never targeted. Third, the axisymmetric RJV may poorly fit storms with asymmetric structure or atypical radial profiles, leaving residual cyclonic perturbations that can redevelop. Together, these limitations identify the primary tuning parameters of the unseeding algorithm: intervention frequency (controlled by the intervention interval and online detection criteria), and vortex-fit fidelity.

Because its parameters can be tuned systematically, vortex manipulation provides a framework for probing environmental limits on TC activity. In seeding experiments, intervention frequency could be increased until additional vortices no longer raise total TC activity because of environmental constraints or spatial packing limits (e.g., Hoogewind et al., 2020). Complementary unseeding experiments could test whether suppressing TCs increases the frequency or intensity of weaker tropical disturbances. In `UNSEED_EX`, for example, the larger number of interventions may reflect either the broader detection criteria or a genuine increase in incipient vortices under the adjusted background state. A `DetectNodes`-only postprocessing analysis could distinguish these possibilities by tracking TD-like waves (e.g., Lin et al., 2025) and other candidate disturbances that do not satisfy the full TC criteria.

The experimental gradation in TC activity produces systematic tropical climate responses, including upper-ocean cooling (Price, 1981; Dare and McBride, 2011) and tropospheric drying (Schenkel and Hart, 2015a; Scoccimarro et al., 2020). Unlike storm-centered composites, these anomalies reflect the climate system's adjustment to sustained changes in TC activity and include shifts in precipitation, cloud amount, and radiative fluxes. Although individual TCs produce local moistening, heavy rainfall, and strong cloud-radiative effects (Schenkel and Hart, 2015a; Hsu et al., 2008; Konrad and Perry, 2010; Bu et al., 2014), greater TC activity here is associated with tropospheric drying, reduced mean precipitation, and weaker cloud radiative forcing. This contrast shows that TC climate effects cannot be represented as a linear superposition of storm anomalies. The TC-masked surface-stress analysis likewise shows that the mean circulation response is dominated by adjustment of the background state rather than direct TC momentum forcing (Sect. 3.3). The broad tropical adjustments found here and by Rios et al. (2026) support viewing TC upscale effects as coupled climate reorganization rather than as independent storm-scale pathways. Online TC modulation is well-suited to this perspective because it alters TC activity while allowing the atmosphere-ocean system to adjust freely.

The framework can also isolate specific pathways through mechanism-denial experiments. Fixed-SST simulations could separate atmospheric drying from SST-mediated effects, while cloud locking could isolate water-vapor and Planck feedbacks from cloud-radiative changes. Process attribution will nevertheless remain limited by parameterized convection and unresolved storm structure until sustained global kilometer-scale simulations become feasible.

Ongoing work is examining the coupled thermodynamic and dynamical adjustments underlying these tropical climate responses. This analysis uses energy and moisture budgets to quantify interactions among TC-induced surface fluxes, convection, radiation, and the large-scale circulation, and to assess whether individual mechanisms such as the TC ocean heat pump can explain the response to persistently altered TC activity. The present study provides the first step in this effort by establishing the experimental framework and documenting the large-scale responses that require process-level explanation.



Code and data availability. The foundational code used to seed and unseed TCs is available at <https://zenodo.org/badge/latestdoi/31787694> (Zarzycki, 2025). The CESM (version 2.3 Alpha 17d) source code to run the AQUA configuration, modifications to the seeding code, output from model experiments, and analysis code can all be found in <https://doi.org/10.26208/BDFC-3E18> (Pan and Zarzycki, 2026). TempestExtremes is located at <https://github.com/ClimateGlobalChange/tempestextremes> (Ullrich et al., 2021).

Appendix A: TempestExtremes TC tracking thresholds

As discussed in Sect. 2.2, the `DetectNodes` binary employs distinct thresholds depending on whether tracking is performed online or offline, as well as between the UNSEED and UNSEED_EX experiments. The specific threshold values for all configurations are compiled in Table A1.

The parameter naming convention reflects the meteorological variable and the geometric constraint applied:

- **Prefixes** denote the tracking variables: `psl` represents Sea Level Pressure (SLP), `wc` represents the upper-tropospheric warm core amplitude (measured by 300–500 hPa thickness), and `rv` represents 850 hPa relative vorticity.
- **Suffixes** define the algorithmic criteria: `Fomag` and `Fodist` indicate the required falloff magnitude and the radial distance (in degrees) over which that falloff must occur. `MaxOffset` specifies the maximum allowable horizontal distance between the secondary variable extremum and the SLP minimum, controlling the degree of vertical vortex tilt permitted.

The `StitchNodes` thresholds used to generate TC trajectories in postprocessing are listed below:

```
TRAJRANGE=6.0
TRAJMINLENGTH=10
TRAJMAXGAP=3
MAXTOPO=150.0
MAXLAT=50.0
MINWIND=10.0
MINLEN=10
```

Examples of `StitchNodes` calls can be found in Ullrich and Zarzycki (2017) and Ullrich et al. (2021), and parameter definitions are detailed in TempestExtremes user documentation.

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Table A1. DetectNodes threshold values for the different tracking configurations.

	Offline	Online SEED(_EX)	Online UNSEED	Online UNSEED_EX
pslFOmag [hPa]	2	4	4	2
pslFOdist [°]	5.5	5.5	5.5	5.5
wcFOmag [m]	-6	-15	-15	-6
wcFOdist [°]	6.5	6.5	6.5	6.5
wcMaxOffset [°]	1	0.5	0.5	0.5
rvFOmag [s ⁻¹]	-8×10^{-5}	N/A	-8×10^{-5}	N/A
rvFOdist [°]	5.5	N/A	5.5	N/A
rvMaxOffset[°]	0.5	N/A	0.5	N/A
maxlat [°]	N/A	30	30	30
mergedist [°]	6	6	6	6

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