# The Unexpected Oceanic Peak in Energy Input to the Atmosphere and its Consequences for Monsoon Rainfall

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**Key Points:**

- The maximum energy input to the atmosphere in boreal summer lies over the northern Indian Ocean
- Cloud radiative effects strongly enhance energy input over ocean compared to land
- Surface heat capacity contrasts interact with cloud radiative effects to shift monsoon rainfall

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**Abstract**

Monsoons have historically been understood to be caused by the low thermal inertia of land, allowing more energy from summer insolation to be transferred to the overlying atmosphere than over adjacent ocean. Here we show that during boreal summer, the global maximum net energy input (NEI) to the atmosphere unexpectedly lies over the Indian Ocean, not over land. Observed radiative fluxes suggest that cloud-radiative effects (CRE) almost double the NEI over ocean, shifting the NEI peak from land to ocean. Global climate model experiments with both land and interactive sea surface temperatures confirm that CRE create the oceanic NEI maximum. Interactions between CRE, NEI, circulation, and land-sea contrast in surface heat capacity shift precipitation from Southeast to South Asia. CRE thus alter the global partitioning of precipitation between land and ocean and the spatial structure of Earth's strongest monsoon, in ways that can be understood through the NEI.

## Plain Language Summary

Land's influence on the energy supplied to the atmosphere has long been recognized as a leading cause of monsoons. From early theories conceptualizing monsoons as continental-scale sea breezes responding to land-sea temperature contrasts, to modern frameworks based on air's total energy content, the energy input to the atmosphere over land has been assumed higher than that over ocean in the summer hemisphere. We show that, instead, in the Asian region, the energy input to the atmosphere is larger over ocean than land because of clouds' effects on radiation. Observations and simulations indicate that the spatial pattern of tropical rainfall is set by interactions between clouds and the land-sea contrast in surface heat capacity, mediated by atmospheric circulation.

## 1 Introduction

Monsoons have, for over a century, been known to be caused by land-sea contrast (Blanford, 1888; Ananthakrishnan et al., 1965). The low thermal inertia of off-equatorial land allows more energy from summer insolation to be transferred to the overlying atmosphere there than over the near-equatorial ocean; this sets up a thermally direct circulation with precipitating ascent over the continent. This precipitating circulation was traditionally seen as a continental-scale sea breeze responding to land-sea temperature contrast, but in recent decades has been better understood by including the latent heat of water vapor in measures of energy, such as the widely used moist static energy (MSE). A general understanding of controls on the structure of monsoons was obtained using a series of idealized climate models in which air's MSE is a central variable (Chou et al., 2001; Neelin, 2007; Plumb, 2007).

As an alternative to theoretical frameworks based on the energy content of air, frameworks based on energy sources, i.e., the net energy input (NEI) to the atmosphere, have been explored (Biasutti et al., 2018). The NEI is the sum of surface turbulent fluxes (sensible and latent heat) and the net radiative flux into the atmospheric column; horizontal contrasts in NEI can be viewed as a forcing for tropical circulations, which are typically "energetically direct" with an ascent branch near the NEI and MSE maxima. Radiative and wind-evaporation feedbacks can render the NEI diagnostic, rather than a true exogenous forcing, but these feedbacks often exhibit substantial cancellation (Peterson & Boos, 2020; Laguë et al., 2021). The seasonal cycle of tropical precipitation maxima is strongly associated with that in NEI and, through conservation of energy, with zonal and meridional energy fluxes carried by time-mean overturning tropical atmospheric circulations (Kang et al., 2008; Donohoe et al., 2013; Boos & Korty, 2016; Adam et al., 2016).

Despite this theoretical focus on NEI as a driver of tropical circulations, few studies have examined observationally-based estimates of NEI, especially with the goal of understanding how observed spatial structures influence regional precipitation. Top-of-atmosphere (TOA) radiative fluxes have been used to show that there is positive NEI over continents in the summer hemisphere, with weaker values over ocean and strong negative NEI over continents in the winter hemisphere (Chou & Neelin, 2003); such studies argued that an energetically direct circulation results, with precipitating ascent over summer continents. A review of the dynamics of tropical convection zones and monsoons (Neelin, 2007) stated that NEI was systematically larger over land than ocean by 50-100 W m$^{-2}$, with that contrast driving planetary-scale monsoon flow. Here we highlight a surprising deviation from this view of land-ocean contrast: an oceanic maximum in NEI that we show strongly influences the spatial structure of precipitation in Asia. We build on prior studies of cloud radiative effects (CRE) in monsoons (Sharma, 1998; Rajeevan & Srinivasan, 2000; J. Li et al., 2017) to show that CRE play a key role in setting this spatial pattern of NEI (Section 4). Using a general circulation model (GCM) that, unlike in prior studies of the influence of CRE on precipitation (Voigt & Albern, 2019; Byrne & Zanna, 2020), accounts for the differing thermal inertia between ocean and land, we show that differences in the response of the land and sea surface to CRE establish this oceanic NEI maximum and set the structure of precipitation (Section 5).

## 2 Materials and Methods

This study uses atmospheric reanalyses, observations, and a global climate model. All data used here are publicly available. Figures 1-2 use the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis Version 5 (ERA5) (Hersbach et al., 2020) (1979-2018) so as to display an internally-consistent estimate of NEI and its components. Findings reported here were verified against other reanalyses and observational products, listed below. Conclusions were based only on features for which all listed datasets displayed qualitative agreement.

### 2.1 Reanalysis Products

In addition to ERA5, we use surface turbulent and radiative fluxes and TOA radiative fluxes from these reanalyses in Figure 1(c):

1. The National Center for Environmental Prediction Climate Forecast System Reanalysis, Version 2 **(CFSR)**(Saha et al., 2014) (1979-2016)
2. The ECMWF Interim Reanalysis **(ERA-I)**(Dee et al., 2011) (1979-2015)
3. The Japanese Meteorological Agency 55-year Reanalysis **(JRA)**(Kobayashi et al., 2015) (1979-2008)
4. The National Aeronautics and Space Administration (NASA) Modern-Era Retrospective Analysis for Research and Applications, version 2 **(MERRA2)**(Gelaro et al., 2017) (1980-2015)
5. The National Center for Environmental Prediction-Department of Energy Reanalysis II **(NCEP)**(Kanamitsu et al., 2002) (1948-2018)

### 2.2 Observational Products

We also use the following observational estimates of ocean surface fluxes, surface and TOA radiative fluxes, cloud fraction, and precipitation:

#### *2.2.1 Air-Sea Turbulent Fluxes*

1. The National Oceanography Centre Surface Flux and Meteorological Dataset **(NOCS)**(Berry & Kent, 2009) (2000-2018)
2. The Woods Hole Oceanographic Institution Objectively-Analyzed Air-Sea Flux Project, version 3 **(OAFlux)**(Yu & Weller, 2007) (1958-2018)

3. The National Oceanic and Atmospheric Administration (NOAA) Climate Data Record of Ocean Heat Fluxes, version 2 **(SeaFlux)**(Clayson et al., 2016) (2000-2020)

#### *2.2.2 Radiative Fluxes*

Clouds and the Earth's Radiant Energy System Energy Balanced and Filled TOA edition-4.0 data product **(CERES)**(Loeb et al., 2018). In figures where this is combined with other datasets, the overlapping years of 2000-2018 are used.

#### *2.2.3 Cloud Fraction*

The Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) General Circulation Model (GCM)-Oriented Cloud CALIPSO Product **(CALIPSO-GOCCP)**(Chepfer et al., 2010) (2001-2018). In Figure S2, we use CALIPSO-GOCCP's definition of high clouds, i.e., clouds above 6.5 km altitude, to calculate high cloud fraction. High cloud fraction is computed as the maximum cloud area fraction over all layers higher than this threshold.

#### *2.2.4 Precipitation*

The Global Precipitation Climatology Project **(GPCP)** version 2.3 (Adler et al., 2018). A climatology was calculated using monthly means from 1979 to 2020.

### 2.3 Estimation of Net Energy Input

We estimate the net energy input (NEI) as the sum of upward surface turbulent fluxes, upward surface radiative fluxes, and downward top-of-atmosphere (TOA) radiative fluxes. All terms needed to calculate this quantity are included in the reanalyses.

In observational products, we use TOA and surface radiative fluxes from CERES with turbulent surface fluxes over ocean from each of the observational products listed. It is difficult to obtain estimates of the global distribution of surface turbulent fluxes over land; however, due to the low heat capacity of land, the net land surface energy flux is near zero on seasonal timescales and therefore the NEI is nearly equal to the TOA flux over land (Neelin, 2007). In some regions, a small amount of energy (generally not exceeding 20 W $m^{-2}$) is consumed at the surface through processes such as seasonal snowmelt; we account for this by computing the difference between NEI and TOA flux over land from ERA5 and applying this as a correction to arrive at the NEI over land in Figure 1(c).

### 2.4 Global Climate Model Experiments

We use the Community Earth System Model (CESM) version 2.0, with the scientifically-validated "ETEST" component set. This consists of a global atmosphere model at a resolution of 2.5°× 1.875°with 32 vertical levels, coupled to a slab ocean 30 m deep, using a climatological q-flux (i.e., a spatially-varying heat flux in the ocean representing the effects of ocean heat transport and processes such as ice melt/freezing) derived from a coupled control run of the model. In this component set, the Community Land Model (CLM5.0) is used with satellite phenology, and greenhouse gas concentrations are held at pre-industrial (year 1850) levels. A 5-year spinup was used before the experiments were performed.

Because we focus on the impacts of cloud-radiative effects during boreal summer, we initiate all experiments from May 1st of the 6th year of a control run. This prevents model drift due to the altered conditions in the experiments from affecting the season of interest. The "noTropicCloud" experiment consists of an ensemble of five simulations in which radiative effects of clouds within the latitudes 35°S-35°N were set to zero, i.e., clouds within the tropical belt were transparent to both shortwave and longwave radiation. This is similar to the method used in the Clouds On-Off Klimate Intercomparison Experiment (COOKIE)(Stevens et al., 2012), except using prognostic instead of prescribed SST. The latitude of 35° was chosen as it corresponds to the latitude where the annual-mean, zonal-mean TOA fluxes change sign. Each of the five simulations was initiated with a different small perturbation.

The control experiment consists of a similarly-designed ensemble, with CRE active. Results presented are averaged over these ensembles. In figures where differences between the control and noTropicCloud are shown, only areas where differences were significant at the 95% level based on a two-tailed t-test are shaded.

## 3 The Observed Distribution of Net Energy Input

During local summer in each hemisphere, NEI is typically largest over land (Figure 1(a, b)), acting as an energy source for the circulation. This pattern is consistent with the view that monsoon circulations are driven by a continental energy source maximum (Neelin, 2007). For South Asia, however, the atmosphere gains substantially more energy over the Bay of Bengal than over adjacent land, which is, according to several datasets, the global maximum of NEI in boreal summer. Despite wide variation in the estimated NEI across reanalyses and observational products (Figure 1(c)), all display an NEI peak over the Northern Indian Ocean.

We decompose the NEI into surface and TOA components, showing that the net surface energy flux (including radiation) is near zero or negative over the Bay of Bengal and Arabian Sea during boreal summer, despite the large surface turbulent heat fluxes into the atmosphere there (Figure 2(a), Figure S1). Over the Northern Indian Ocean, TOA fluxes contribute most of the positive NEI (Figure 2(b)), suggesting a role played by processes that influence TOA radiation, such as clouds. The shortwave and longwave components of the CRE (Figure 2(c)) confirm this: while the shortwave effect of clouds reflects energy into space and is therefore negative over the region experiencing monsoon rainfall, the longwave effect, which retains energy in the atmospheric column, is largest over the Bay of Bengal NEI maximum. This reduction in energy loss to space coincides with an area covered by high cloud tops (Figure S2); the frequent occurrence of organized mesoscale convective systems in this region likely contributes to this large high-cloud fraction (P.-J. Chen et al., 2021; Hamada et al., 2014; Yuan & Houze, 2010; Luo et al., 2017). The resulting net CRE (Figure 2(d)) thus makes a large positive contribution to the NEI over the northern Indian Ocean.

## 4 The Prognostic Influence of Cloud Radiative Effects

While observed radiative fluxes can be used to estimate the net influence of clouds on radiation given the observed atmospheric state (e.g. Figure 2(d)), it is possible that large changes in wind, temperature, humidity, and cloud properties would occur in the absence of CRE. This motivates our use of the climate model described in Section 2 to determine, prognostically, the influence of CRE on both the NEI and the large-scale circulation.

The control run captures key features of the NEI distribution, including the energy sources over ocean in the Southern Hemisphere and Northern Hemisphere continents, and the maximum over the northern Indian Ocean during boreal summer (Figure 3(a)). There is some bias relative to ERA5, but this is of comparable magnitude to the observational uncertainty in NEI (e.g. Fig. 1(c)). The CRE contribution to NEI (Figure 3(b)), calculated as the difference between clear-sky and all-sky radiative effects, in the control run is similar to that in observations (Figure 2(d)).

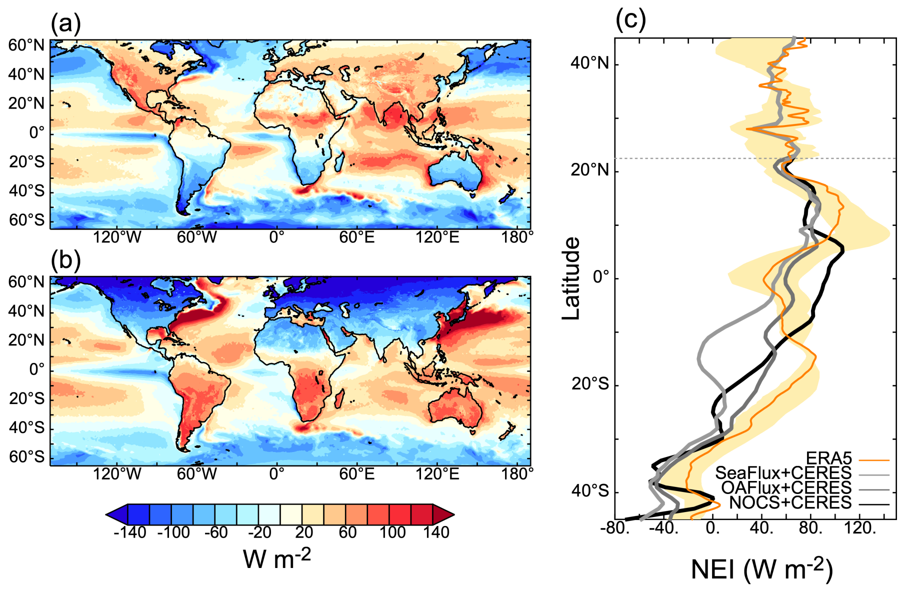


**Figure 1.** The oceanic nature of the energy input maximum during boreal summer: The climatological net energy input (W/m$^2$) to the atmospheric column in (a) boreal summer (June-August) and (b) austral summer (December-February) from ERA5. (c) Net energy input into the atmosphere from three observational estimates (thick grey lines) and ERA5 (orange line) averaged over the longitudes of the Bay of Bengal (90°E-95°E) in boreal summer. The filled area indicates the range of the same quantity from five other reanalysis products (listed in Section 2.1). The dotted line indicates the latitude of the northern edge of the Bay of Bengal.

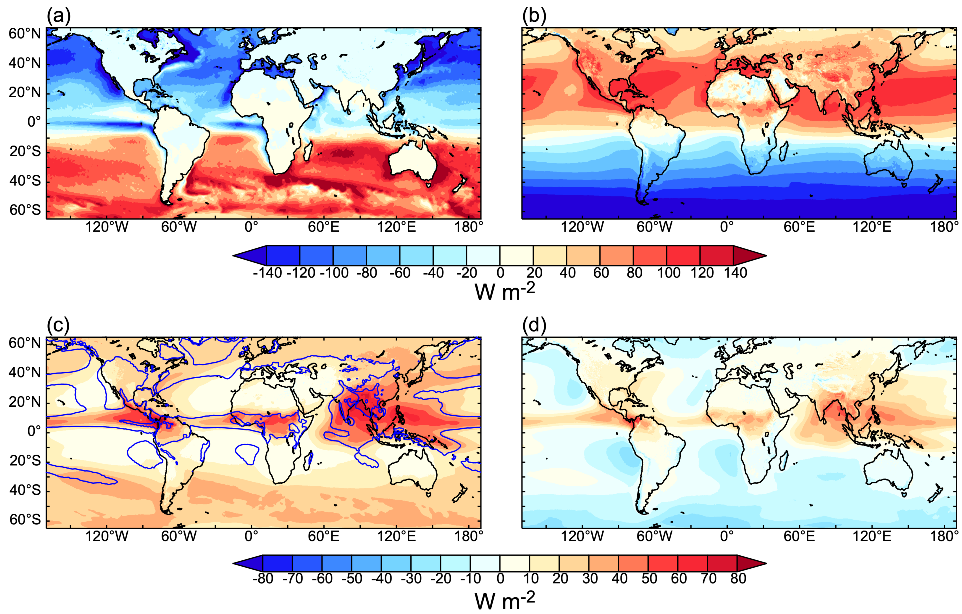


**Figure 2.** Components of the observed energy input: The contributions of fluxes at (a) the surface and (b) the top of the atmosphere to the climatological NEI. Panel (c) shows the contributions of the longwave (colors) and shortwave (contours; intervals of -50 $\mathrm{Wm}^{-2}$) components of CRE respectively. The total estimated contribution of CRE is shown in (d). Quantities are positive if they contribute to the energy content of the atmospheric column.

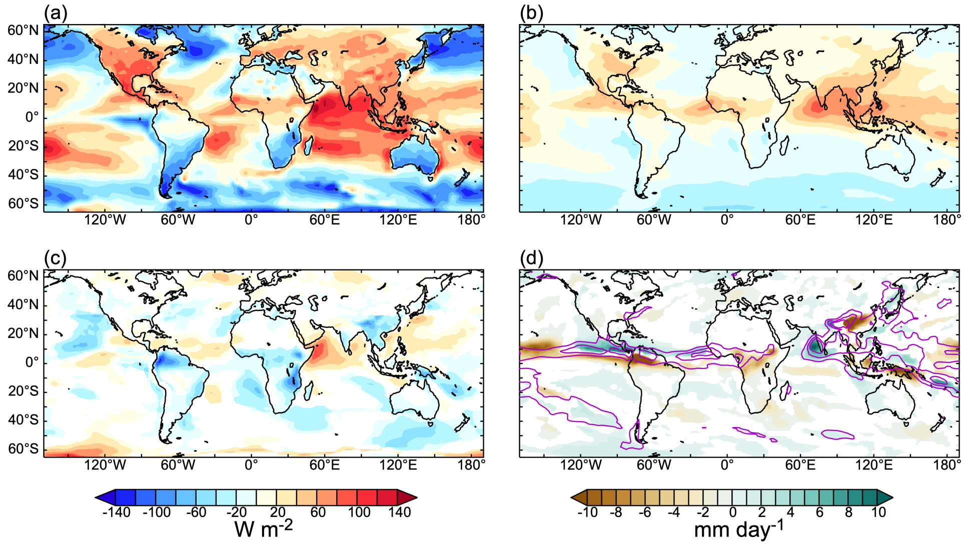


**Figure 3.** Perturbing cloud-radiative effects in a global model: The June-August mean (a) NEI, and (b) CRE contribution to NEI from the control ensemble, calculated as the difference between all-sky and clear-sky radiative effects. The bottom row shows the June-August mean contribution of CRE inferred prognostically (control minus noTropicCloud) (c) to NEI and (d) to precipitation. Contours in (d) indicate the June-August mean precipitation in the control ensemble at 5-mm/day increments.

Examining the difference between the control run and the run with CRE eliminated in the tropics (the noTropicCloud experiment), confirms that clouds enhance NEI over the entire Northern Indian Ocean (Figure 3(c)). However, tropical CRE also reduces NEI over several land areas, particularly Southeast Asia. Overall, the response to removing clouds in the model experiment differs greatly from the CRE inferred simply as the difference between all-sky and clear-sky radiative effects (Figure 3(b)), indicating that CRE induces feedbacks on surface turbulent fluxes and radiation.

The absence of CRE substantially alters boreal summer precipitation (Figure 3(d)). Notably, including CRE reduces rainfall over tropical land relative to that over ocean. The fraction of total summer rainfall within the latitude range of eliminated CRE (35°S-35°N) that occurs over land increases from 0.2 in the control to 0.25 in the noTropicCloud experiment (a relative increase of 26% (±4%, one standard deviation)); in the deep tropics (20°S-20°N), this re-partitioning of rainfall over land versus ocean is even more pronounced, with a relative increase of 31% (±7%).

The spatial pattern of changes to rainfall is dominated by shifts in precipitation maxima. Over the Atlantic and East Pacific, the oceanic ITCZ is displaced to the north when CRE are included. This is consistent with previous aquaplanet studies (Voigt et al., 2014) that concluded that CRE shift the ITCZ poleward by producing interhemispheric NEI asymmetries. Over the Indo-Pacific, however, the spatial pattern of changes in precipitation is more complex, displaying a striking southwestward shift of rainfall from East Asia to South Asia when CRE is added.

## 5 The Influence of Cloud Radiative Effects on Atmospheric Circulation

The cause of this mostly zonal shift over Asia can be understood using dynamic or energetic perspectives. We first describe how CRE alter the distributions of precipitation, surface enthalpy fluxes, and MSE advection, then use an energetic framework to show how the influence of clouds on NEI is quantitatively consistent with the model-simulated precipitation shift.

We treat the noTropicCloud experiment as a basic state on which CRE can be applied. In that state, boreal summer precipitation peaks over southeastern Asia (Figure 4(a), gray contours), consistent with idealized model simulations that show monsoon precipitation concentrates over the eastern part of rectangular tropical continents due to the Rossby gyres that comprise three-dimensional monsoon circulations (Chou et al., 2001; Privé & Plumb, 2007; S.-P. Xie et al., 1999). In those studies, much of this concentration of rainfall over the eastern part of the continent is due to advection of dry air by the lower-tropospheric Rossby gyre. We see evidence for that in the noTropicCloud run: the strong zonal MSE gradient over South Asia (Figure 4(a)) is spanned by low-level eastward winds that feed into the region of peak precipitating ascent (Figure 4(b)), as expected for the linear Rossby gyre component of a monsoon (Gill, 1980; Hoskins & Rodwell, 1995). The resulting advection of MSE, vertically integrated over the atmosphere, provides a negative energy tendency over much of South and Southwest Asia exceeding 150 W m$^{-2}$ (Figure 4(b)). Horizontal advection by the Rossby gyre thus greatly compensates the radiative forcing for precipitation over South Asia in the absence of CRE.

With tropical CRE turned on, the shortwave effects of clouds over southeastern Asia reduce surface enthalpy fluxes there by about 100 W m$^{-2}$ (Figure 4(c)). Although the longwave effects warm the atmosphere by 30-50 W m$^{-2}$, opposing the shortwave contribution to the NEI, the net CRE is negative, weakening the thermally-forced Rossby gyre. CRE also convectively stabilize the troposphere, as evidenced by the upper-tropospheric warming and lower-tropospheric cooling seen in the response to tropical CRE (Figure S3). This convective stabilization over land leads to a reduction in precipitating ascent over Southeast Asia (Figure 4(e)), a weakening of the low-level eastward inflow to that region, and a reduction in the negative MSE advection over northern India accomplished by that inflow (Figure 4(d)). This reduction in negative MSE advection peaks around 200 W m$^{-2}$, and is accompanied by enhanced low-level MSE over South Asia and increased precipitation there (Figure 4(f)); note the MSE increase over Southeast Asia peaks in the mid-troposphere, consistent with its modification by free-tropospheric CRE rather than low-level moisture advection. In summary, CRE convectively stabilizes the continental precipitation maximum and weakens the associated Rossby gyre, reducing the dry air advection that would otherwise suppress precipitating ascent over South Asia (Figure 4(e)).

One can alternatively view this process in terms of the influence of CRE on NEI over land, which is negative because the shortwave part of CRE exceeds the longwave part there. A negative NEI anomaly is thus induced over Southeast Asia by CRE, and this must be balanced by an anomalous flux of energy into the region, which in the tropics is typically accomplished by time-mean overturning circulations (Kang et al., 2008; Boos & Korty, 2016). Figure 4(g) shows the energy flux prime meridian (EFPM) in the control and noTropicCloud experiment, with the EFPM being the zero line of the divergent eastward energy flux (vertically integrated over the atmosphere); the EFPM is expected to move together with zonal shifts in zonal overturning circulations (Boos & Korty, 2016), similar to the way the energy flux equator (EFE) moves with meridional shifts in meridional overturning circulations (Kang et al., 2008). The inclusion of CRE, by altering the spatial pattern of NEI, shifts the EFPM westward by 5.8°, closely matching the location of the EFPM in reanalyses over the Bay of Bengal (Boos & Korty, 2016).

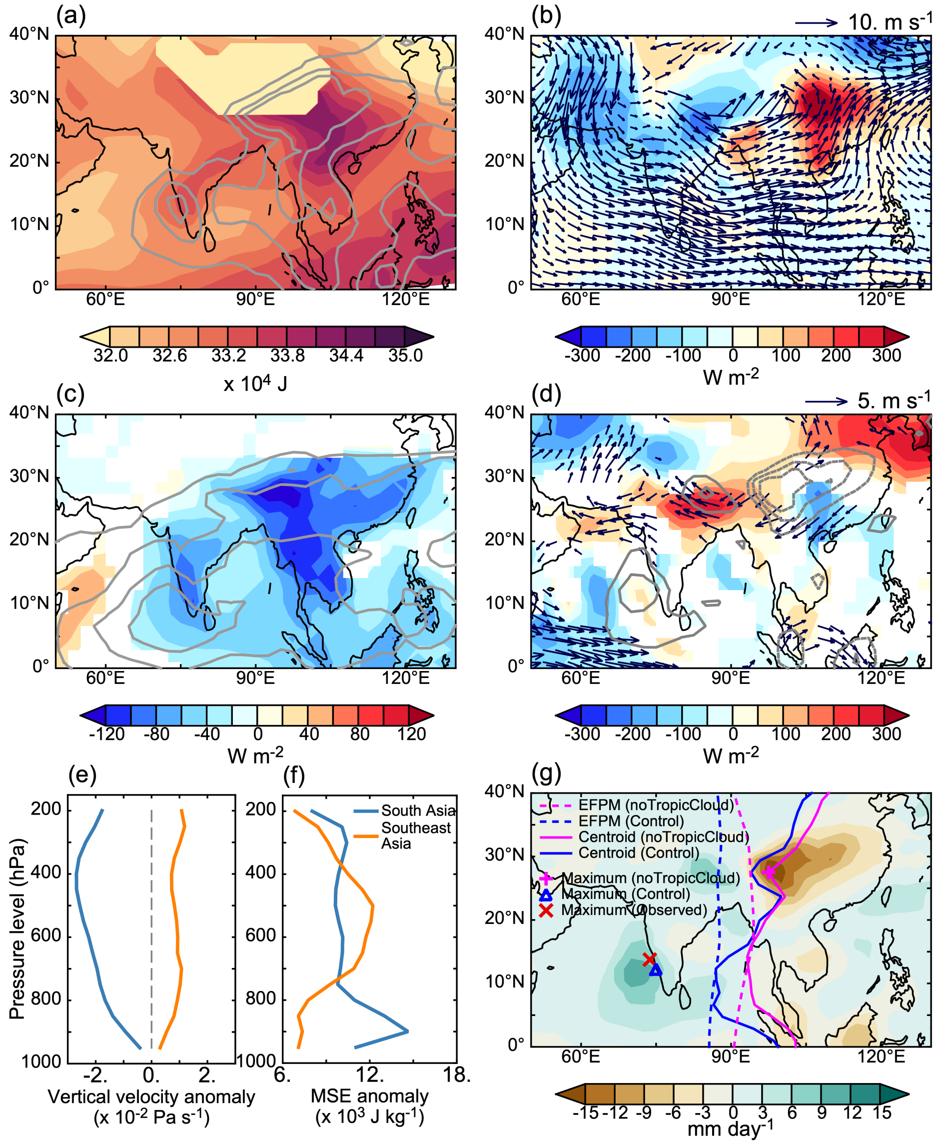


**Figure 4.** CRE induce a westward shift and weakening of the precipitating Rossby gyre over Asia: (a) Precipitation (grey contours, interval 5 mm day$^{-1}$) and MSE at 700 hPa (shading) in the noTropicCloud experiment. (b) Vertically-integrated MSE advection in the absence of CRE (shading) with wind velocity (arrows) at 700 hPa. (c) The anomaly (control minus NoTropicCloud) in surface turbulent fluxes (shading, W m$^{-2}$) and column-integrated radiative flux convergence (grey contours, interval 20 W m$^{-2}$, negative contours dashed and zero contour omitted). (d) Anomaly (control minus NoTropicCloud) in the quantities shown in (b) and precipitation (grey contours, interval 5 mm day$^{-1}$, negative contours dashed and zero contour omitted). (e) and (f) show the anomaly (control minus noTropicCloud) in vertical velocity and MSE, respectively, averaged over South (70°E-90°E, 10°N-30°N) and Southeast (90°E-110°E, 10°N-30°N) Asia. (g) Anomaly (control minus noTropicCloud) in precipitation (shading) with the EFPM (dashed lines) and precipitation centroid (solid lines). Symbols indicate the location of maximum seasonal-mean precipitation.

A corresponding westward shift in precipitation occurs, with the precipitation centroid moving 4.3°westward in the meridional mean over the region shown. This constitutes good agreement, as shifts in the zero lines of divergent energy flux are typically highly correlated with but larger than the shifts in precipitation maxima (Kang et al., 2008; Shekhar & Boos, 2016). Including CRE also shifts the precipitation maximum from continental Southeast Asia to its observed location over ocean (Figure 4(g)).

## 6 Discussion

Our analysis of the observed NEI distribution revealed that in boreal summer, the global maximum NEI is positioned over the northern Indian Ocean rather than over land, challenging the conventional view that large-scale tropical circulations in solstice seasons are associated with continental NEI maxima. When the NEI was decomposed, CRE were found to be the primary contributor to this maximum. This is distinct from other observed NEI maxima over oceans, where turbulent surface fluxes dominate (e.g., the western boundary currents and trade wind regions in the winter hemisphere; Figure 1(a,b)).

Prior studies have examined the distribution of CRE in monsoons, showing, e.g., that the observed net CRE in the Asian monsoon is negative (Rajeevan & Srinivasan, 2000), and that net CRE over Asia is more negative for higher-altitude cloud tops (Saud et al., 2016). In simulations with realistic boundary conditions, CRE have been shown to amplify natural modes of Asian monsoon variability (Lu et al., 2021). Previous aquaplanet studies identified meridional shifts of precipitation maxima in response to CRE (Voigt et al., 2014; Randall et al., 1989; Byrne & Zanna, 2020; Harrop & Hartmann, 2016; Popp & Silvers, 2017); in contrast, we found that with realistic continents, the primary response to CRE over Asia is instead a zonal shift. This zonal shift is produced by the contrasting effects of CRE over land and ocean combined with the three-dimensional large-scale tropical circulation. Over land, low surface heat capacity allows the shortwave effect of clouds to cool the surface and convectively stabilize the atmosphere; over ocean, shortwave CRE has a weaker effect due to the ocean's high heat capacity. This means that over ocean, longwave CRE is the dominant contributor to the NEI, warming the atmosphere (Randall et al., 1989) even though shortwave and longwave CRE approximately cancel at TOA (Tian & Ramanathan, 2002).

The asymmetry in CRE between ocean and land is reflected in the large increase in the proportion of rainfall occurring over land when CRE is eliminated (a relative increase of 26%) and the inland shift of the location of maximum precipitation (Figure 4(g)). When the NEI and convective instability are reduced in the region of the precipitation maximum, the Rossby gyre circulation weakens, allowing precipitation to shift westward. This reduction in convective activity over land is consistent with theoretical models showing that CRE provides a negative feedback on the response to forcings over land (Zeng & Neelin, 1999), in contrast with the positive feedback on circulations that CRE can provide over ocean (Su & Neelin, 2002). The dry and wet biases in CESM and many other climate models over continental South (Sperber et al., 2013; S. Xie et al., 2012) and Southeast Asia (Ma et al., 2014; W.-T. Chen et al., 2019), respectively, suggest that the true magnitude of this response may be larger than that seen in our experiments.

Our findings prompt a rethinking of the role of land-sea contrast in setting the distribution of tropical NEI: in the largest monsoon system, the NEI maximum lies over ocean instead of land. Our findings also highlight the importance of differences between the land and ocean response to CRE. While CRE have long been recognized as a crucial process in atmospheric circulation (Randall et al., 1989; Tian & Ramanathan, 2002; Sherwood et al., 1994; Sohn & Smith, 1992; Slingo & Slingo, 1988; Y. Li et al., 2015) and a key determinant of its response to increasing greenhouse gas concentrations (Hansen et al., 1984; Voigt & Shaw, 2015; Voigt & Albern, 2019; Ceppi & Hartmann, 2016; Ceppi et al., 2017), they have frequently been studied in aquaplanets (Voigt et al., 2014; Randall et al., 1989; Byrne & Zanna, 2020; Harrop & Hartmann, 2016; Voigt & Shaw, 2015; Ceppi & Hartmann, 2016) or settings where the ocean's heat capacity is unaccounted for (Sherwood et al., 1994; Slingo & Slingo, 1988; Y. Li et al., 2015; Hansen et al., 1984). The results of this study suggest that the way forward in understanding the impacts of CRE on atmospheric circulation and patterns of precipitation must necessarily include the effects of spatial contrasts in the heat capacity of the underlying surface.

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## Open Research and Data Availability Statement

All observational and reanalysis data used in this study are publicly available from the sources listed below:

- ERA5 (Hersbach et al., 2020): `https://confluence.ecmwf.int/display/CKB/How+to+download+ERA5`
- CFSR (Saha et al., 2014): `https://www.ncei.noaa.gov/access/metadata/landing-page/bin/iso?id=gov.noaa.ncdc:C00765`
- JRA (Kobayashi et al., 2015): `https://jra.kishou.go.jp/JRA-55/index_en.html`
- MERRA2 (Gelaro et al., 2017): `https://disc.gsfc.nasa.gov/datasets?sort=-timeRes&project=MERRA-2`
- NCEP (Kanamitsu et al., 2002): `https://psl.noaa.gov/data/gridded/data.ncep.reanalysis2.html`
- NOCS (Berry & Kent, 2009): `https://noc.ac.uk/science/sustained-observations/noc-surface-flux-dataset`
- OAFlux (Yu & Weller, 2007): `http://apdrc.soest.hawaii.edu/datadoc/whoi_oafluxmon.php`
- SeaFlux (Clayson et al., 2016): `https://seaflux.org/data-2/data`
- CERES (Loeb et al., 2018): `https://ceres.larc.nasa.gov/data/`
- CALIPSO-GOCCP (Chepfer et al., 2010): `https://climserv.ipsl.polytechnique.fr/cfmip-obs/`
- GPCP (Adler et al., 2018): `http://eagle1.umd.edu/GPCP_ICDR/`

The output of the described climate model experiments, along with instructions to reproduce the experiments, are available in a Zenodo repository (DOI:10.5281/zenodo.5704060).

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# Supporting Information for "The Unexpected Oceanic Peak in Energy Input to the Atmosphere and its Consequences for Monsoon Rainfall"

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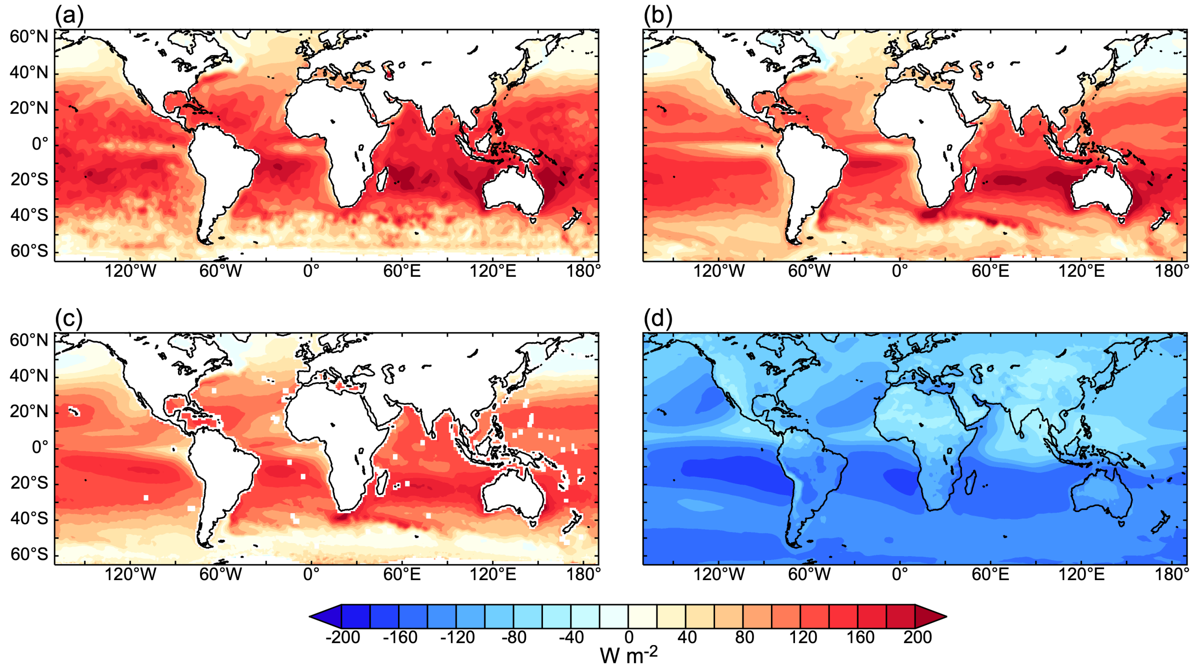


**Figure S1.** Climatological turbulent air-sea fluxes into the atmosphere during June-August from various data sources (a-c) and net radiative flux convergence from CERES vertically integrated over the atmosphere (d): (a) NOCS, (b) OAFlux, (c) SeaFlux. Positive values indicate a flux into the atmospheric column.

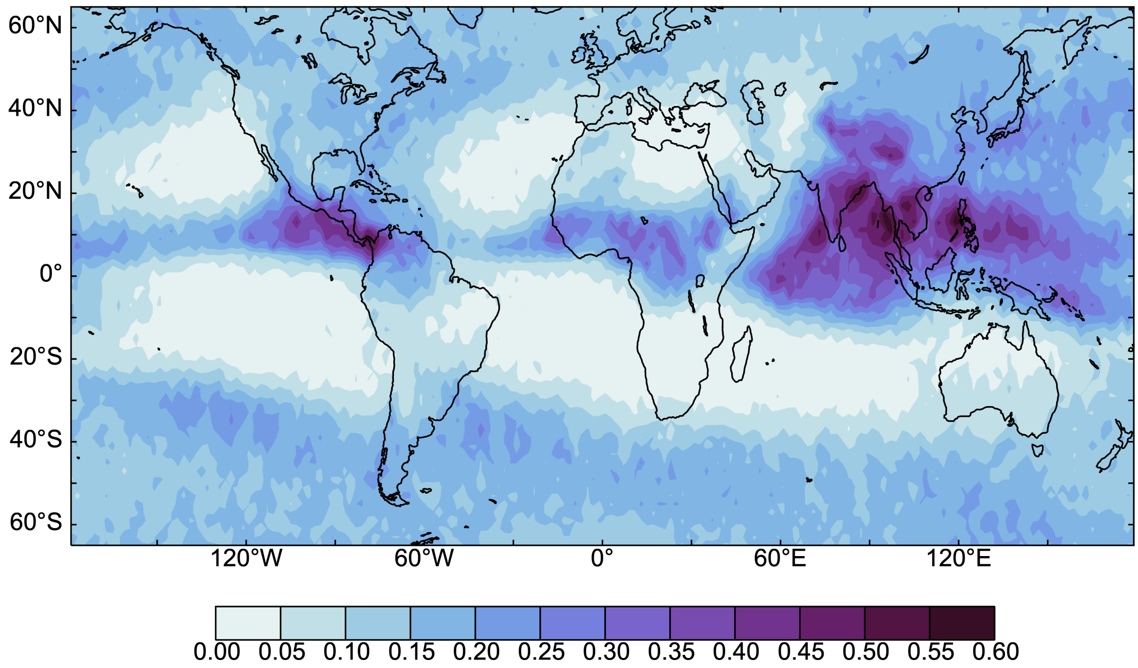


**Figure S2.** The climatological high cloud fraction during June-August from CALIPSO-GOCCP. Note the large high cloud fraction over the South Asian region and in particular, the Bay of Bengal.

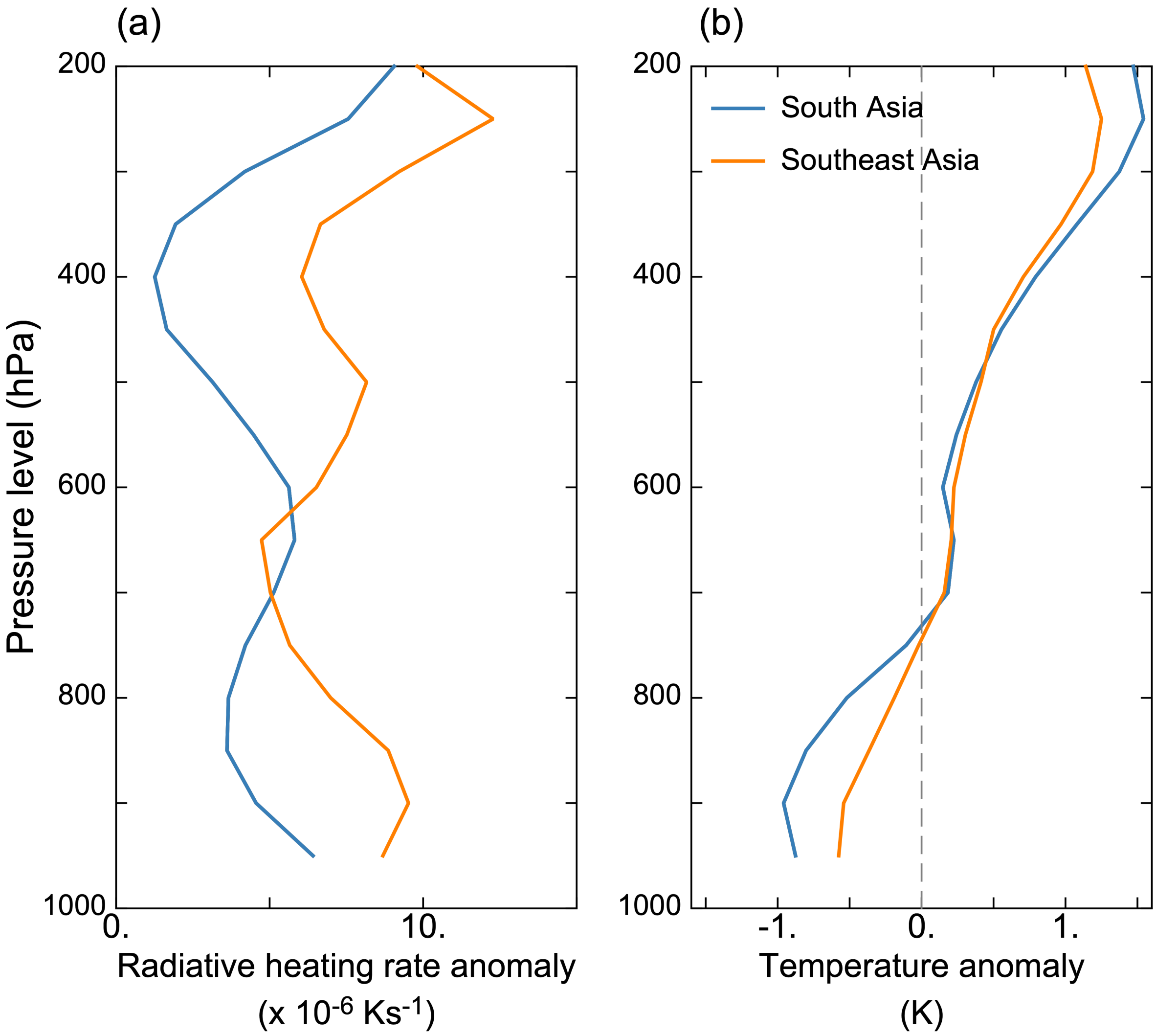


**Figure S3.** The anomaly (control minus noTropicCloud) in (a) radiative heating rate and (b) temperature averaged over South (70°E-90°E, 10°N-30°N) and Southeast (90°E-110°E, 10°N-30°N) Asia. The net effect of CRE on the tropospheric radiative heating in both locations is positive, and is accompanied by a large, O(100 W m$^{-2}$), reduction in surface fluxes of sensible and latent heat; these effects together produce the convective stabilization of the atmosphere seen in the anomalous temperature profiles in (b).