

# Greater North American Monsoon Precipitation Brought by Enhanced Storm Activity in a Warm Climate

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## Research Article

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# Greater North American Monsoon Precipitation Brought by Enhanced Storm Activity in a Warm Climate

## Abstract

Climate models suggest a weakened North American summer monsoon (NAM) under elevated atmospheric CO<sub>2</sub>, posing serious challenges to NAM region's socioeconomic development. This outlook is consistent with simulations of the mid-Pliocene (3.0 - 3.3 Ma), the closest analog to the middle-of-the-road future scenario. However, proxy evidence reveals enhanced summer precipitation in the monsoon region during the mid-Pliocene, raising questions about the climate projections. Here, we develop novel coupled high-resolution atmosphere-ocean simulations and utilize a feature tracking algorithm for mesoscale convective systems. Using these tools, we show that this summer precipitation enhancement reflects enhanced mesoscale convective system activity during the mid-Pliocene—an effect that emerges only in high-resolution simulations and results from land surface greening. These results highlight the importance of resolving surface albedo-moist convection interactions in modulating hydroclimate conditions in semi-arid continents, where the skill of climate hindcasts and projections are likely limited by model resolution.

**Keywords:** North American Monsoon, Mesoscale Convective Systems, Mid-Pliocene Warm Period, Hydroclimate Variability

## 1 Introduction

In modern climate, precipitation from Mexico to southern California is dominated by the North American Monsoon (NAM). NAM precipitation is projected to decrease in the future due to the weakening of the NAM circulation as a result of enhanced sea surface temperature (SST) contrasts between the tropics and the extratropics along the eastern tropical Pacific [1, 2]. However, most projections come from global climate models at horizontal grid spacings of 100 km or coarser (here referred to as low resolution, LR) [2, 3]. LR simulations potentially lack representation of important characteristics of NAM precipitation, its dependency on topography, and Gulf of California moisture influx [4–13]. Most notably, LR simulations cannot represent the mesoscale convective systems (MCSs) [14] that are a key feature of NAM precipitation

[15, 16]. This limitation, among others, casts doubt on the fidelity of LR simulations to predict responses of NAM precipitation to external climate forcings.

To assess model skill at simulating NAM responses to external forcings, studies have looked into the Mid-Pliocene warm interval (3.0 - 3.3 Million years ago, Ma) [17, 18]. It is widely recognized that the mid-Pliocene provides a useful benchmark for understanding climate system behavior under sustained global warmth. It featured similar topography, geography, and biome assemblages to today as well as global mean temperatures comparable to those projected for "middle-of-the-road" 21st century scenarios (2 - 4 °C above preindustrial) [19, 20]. While certain aspects of the mid-Pliocene climate system, such as Atlantic overturning strength, ENSO variability, and vegetation distributions, differ from those expected under future high-CO<sub>2</sub> conditions [21–24], the interval nonetheless provides a well-constrained and physically relevant test bed for examining climate responses under sustained global warmth. Accordingly, we treat the mid-Pliocene as a robust warm-climate benchmark rather than as a strict one-to-one analog for any specific future scenario. To study this interval, the paleoclimate community has conducted several generations of the Pliocene Model Intercomparison Project (PlioMIP) [25, 26]. The average across the most recent PlioMIP2 simulations, all of which are low-resolution, shows a similar reduction in NAM precipitation with warming [2, 18, 26] (Supplementary Fig. 1) that is consistent with predicted future changes by LR models.

In contrast, proxy records from the mid-Pliocene do not support reduction in NAM precipitation during this warm interval. Fossil assemblages from Baja California include multiple species of freshwater *Crocodylus*, which require persistent aquatic habitats [27]. This implies that wetter conditions than today would have been present because modern Baja is too arid to sustain such habitats. Additional faunal, floral, and geomorphic evidence from the NAM region, as summarized by Bhattacharya et al. [18], further supports this interpretation, highlighting agreement across multiple independent proxies. Isotopic evidence from DSDP Site 475 near the core modern NAM domain (23.05°N, 109.05°W) ([18]; Fig. 1, e and f) likewise indicates isotopically enriched precipitation, suggesting an intensification of Pliocene summer convective rainfall [18]. These multiple lines of evidence therefore point to greater precipitation, especially greater summer precipitation, in the NAM region during the mid-Pliocene, highlighting a key discrepancy with LR model hindcasts. This long-standing proxy-model discrepancy raises the question of whether coarse model resolution, rather than boundary forcing alone, limits the ability of simulations to reproduce the wetter mid-Pliocene NAM indicated by proxies. Thus, in this study, we investigate whether an atmosphere-ocean coupled model with a high-resolution (HR; tens of km grid spacing) atmosphere can simulate enhanced NAM precipitation to mid-Pliocene climate forcings. At this resolution, the model can represent the mesoscale organization of MCSs, the main component of NAM precipitation, which are otherwise aliased into other forms of precipitation in LR simulations.

A defining characteristic of NAM rainfall is its pronounced diurnal cycle of moisture flux and storms over the Sierra Madre Occidental (SMO, 1a). During the day, moisture converges towards the SMO from the Gulf of California and the coastal Pacific,

77 bringing thunderstorms and precipitation in the late afternoon and early evening. Pre-  
78 cipitation often continues through the night, at which point the pattern reverses and  
79 moisture divergence is observed along the slopes of the SMO. Storms dissipate by late  
80 night or early morning, and the atmosphere over land stabilizes [28, 29]. This diurnal  
81 cycle provides the environment in which MCSs organize, making them a dominant  
82 contributor to NAM rainfall [4, 5, 30].

83 Distinctive from scattered or isolated convection, MCSs are large clusters of cumu-  
84 lonimbus clouds, which can span up to several hundred kilometers, last more than 24  
85 hours, and consist of active convective towers and broad bands of stratiform regions  
86 [31]. In the core NAM domain, MCSs contribute upwards of 50-60% of summer pre-  
87 cipitation [30]. Enhancing climate model resolution has been shown to substantially  
88 improve the simulation of Gulf of California moisture transport and NAM precipitation  
89 [11, 12], yet, the contributions from MCSs have not been fully studied.

90 MCSs cannot be explicitly resolved in LR simulations. At low resolutions, ( $\geq 100$   
91 km), key features of MCSs, such as their tilted vertical flow structure, associated  
92 heating profile, and possible meso-cyclone core, are not represented because their  
93 intrinsic scales are finer than the model grid spacing. As a result, most MCSs become  
94 indistinguishable from parameterized scattered moist convection within a grid cell.  
95 Correspondingly, coarse-resolution models tend to substantially underestimate both  
96 the frequency of MCSs and the magnitude of associated extreme precipitation in  
97 regions with frequent MCS activity [32]. At a tens of km grid spacing, mesoscale  
98 circulation dynamics can be partly resolved in combination with parameterized cumu-  
99 lonimbus, e.g. [14]. This hybrid model behavior allows simulations of the propagation  
100 and vertical heating profile of MCSs, albeit with biases, e.g. [14, 33]. Nonetheless,  
101 simulated spatial distribution and MCS statistics are reasonable [34–36].

102 Understanding how MCSs may change in a warm climate such as the mid-Pliocene  
103 is key to mitigating hazards due to climate change for local communities. Summer  
104 MCSs have been associated with an increase in lightning-ignited fires [37] and signifi-  
105 cant flooding [38]. Although outside the NAM domain, convection-permitting (3-4 km)  
106 regional simulations over the central U.S. predict a significant increase in summertime  
107 MCS frequency [39, 40] and its associated precipitation [41] by end-of-century under  
108 RCP8.5. Complementing these findings, regional simulations reveal notable changes  
109 in future projections of NAM behavior at finer scales—including extended monsoon  
110 duration and increases in extreme rainfall, particularly for northern Mexico [42, 43].  
111 These changes are inconsistent with most LR global simulations, including CMIP5 and  
112 CMIP6 models, which generally project a weakened NAM and reduced monsoon pre-  
113 cipitation for end-of-century [44, 45]. To this end, our investigation into a past warm  
114 interval with an high-resolution global model will not only shed lights on the diverse  
115 projections of hydroclimate conditions in this region, but also reveal key dynamical  
116 processes that may operate differently under different low and high spatial resolutions  
117 in a climate model.

## 2 Results

### 2.1 Mean-state differences and mid-Pliocene NAM precipitation changes in high- versus low-resolution simulations

We carried out a set of high- and low-resolution simulations using the Community Earth System Model version 1.3 (CESM1.3) (Materials and Methods) to investigate whether the HR configuration better simulates enhanced mid-Pliocene summer precipitation in the NAM region as suggested by proxy records [18]. These high- and low-resolution simulations separately feature preindustrial (HR-PI and LR-PI) and mid-Pliocene (HR-MP and LR-MP) boundary and forcing conditions. All simulations use the same  $\sim 100$  km horizontal grid spacing ocean and sea ice model. The atmosphere and land grid spacing are  $\sim 25$  km for the HR and  $\sim 100$  km for the LR simulations. Comparison between HR and LR simulations can isolate the effects of enhanced atmosphere and land resolution on model skill because of the same ocean and sea ice components. Our mid-Pliocene boundary conditions include changes to topography, coastal lines, vegetation, and ice sheets that were derived from PRISM4 reconstructions [46]. Following the recommendation of PlioMIP2 [26], the  $\text{CO}_2$  level for the mid-Pliocene is set at 400 ppm, consistent with the majority of  $\text{CO}_2$  reconstructions [47].

Consistent with published studies, e.g. [11, 13, 48], HR-PI shows higher fidelity of simulating NAM precipitation features compared to LR-PI. In HR-PI, the SMO is more narrowly defined along the west coast of Mexico and sits at an elevation 660 meters higher than LR-PI. (Fig. 1, a and b). Subsequently, the core monsoon band, shown by the June–September mean precipitation, is closely anchored to the western windward side of the SMO (Fig. 1d). The simulated core monsoon band in HR-PI is comparable to the core monsoon band seen in a  $0.25^\circ \times 0.25^\circ$  30-year (1991–2020) average of ERA5, a present-day atmospheric reanalysis from ECMWF [49] data, and in a  $0.1^\circ \times 0.1^\circ$  24-year (2001–2024) average of the NASA satellite product IMERG [50] (Supplementary Fig. 2). In contrast, a more diffusive band of precipitation runs across the SMO in the LR simulation (Fig. 1c) due to the less pronounced rain shadow associated with the poorly resolved SMO elevation (Fig. 1, c and d). Additionally, the windward rain band extends farther north in the HR-PI, with more pronounced westerlies from the Pacific and an advection of moisture from the Gulf of California. The former is consistent with greater mechanical forcing [13]. The latter can be explained by the Gulf of California being better resolved in the HR simulation but not in the LR simulation (Fig. 1, a and b) [11, 12].

Although enhanced resolution is well-known to improve model skill at simulating NAM, little is known about its impact on forced responses. In our experiments, despite giving the same mid-Pliocene climate forcing conditions, including topographic changes, HR configuration produced distinct precipitation responses over the SMO and along the Gulf of Mexico from LR configuration. In both regions, HR-MP simulates increased precipitation throughout the NAM season (June–September; Supplementary Fig. 3) compared to HR-PI. This increase is especially notable during the early

summer (June-July). In contrast, LR-MP simulates decreased precipitation throughout the NAM season compared to LR-PI (Fig. 1; Supplementary Fig. 4). For the rest of the paper, we will focus on the early-summer (JJ) results to further explore this difference.

The simulated enhanced summer precipitation over the SMO by HR-MP represents a key improvement in data-model comparison compared to previous global climate model results and our LR-MP simulation, all of which tend to simulate uniform precipitation reduction over this region [18] (Supplementary Fig. 1, 4). Increased precipitation over the SMO may contribute additional runoff to the downstream river basins following the river network (Fig. 1e, f). This net input of runoff is expected to push basin-wide water balance among precipitation, evapotranspiration, and runoff towards positive, which favors mesic environmental conditions [18]. In addition, at DSDP 475, the deuterium composition of terrestrial leaf waxes may reflect material transported from high-elevation catchments via rivers and dust in addition to coastal plains. Hence, the enriched values [18] are also consistent with the increased precipitation over the SMO (Fig. 1f).

## 2.2 Increased NAM precipitation due to enhanced mesoscale convective system activities

To identify key processes that allow HR-MP to simulate enhanced precipitation over the SMO, which is absent in LR-MP, we first compare simulated mean regional circulation responses to mid-Pliocene climate forcings between these two resolutions. Relative to their PI controls, both experiments featured an anticyclonic circulation anomaly to the west of the SMO (Fig. 1, e and f). This anomaly weakens NAM precipitation by redirecting the onshore westerly monsoon flow southward and by advecting dry air from the subtropical high toward the Mexican coast. This anticyclonic circulation anomaly is found to be a key component of NAM precipitation weakening due to forcings including the lowered elevation of the Sierra Madres and the Mexican Plateau [13] or enhanced tropical-subtropical sea surface temperature (SST) gradient in the eastern Pacific [1]. In both HR-MP and LR-MP, the elevation of this region has been lowered by the same amount ( $\sim 300$  m; Supplementary Fig. 5) due to the isostatic adjustment to the loss of Greenland and western Antarctic ice sheets. Simulated SST response patterns also show an enhanced tropical-subtropical gradient in the eastern Pacific (Supplementary Fig. 6). Both changes are conducive to the generation of anticyclonic circulation along the Pacific coast that weakens the onshore monsoon winds, which is counterproductive to precipitation over the northern Mexico. In fact, simulated precipitation reduction along coastal northern Mexico in both LR-MP and HR-MP is consistent with this circulation response. However, this circulation change clearly does not explain the enhanced summer precipitation over the SMO, which occurs only in HR-MP. We subsequently focus on differences in simulated weather features to explore this key difference between HR-MP and LR-MP.

Given the importance of MCS in generating summer precipitation in this region, we implemented a new MCS tracking methodology using the TempestExtremes feature-tracking framework [51]. While TempestExtremes provides generalized detection and tracking algorithms, identifying MCSs requires phenomenon-specific parameter

choices; therefore, detection variables and algorithm parameters were calibrated using simulated late-20th century (1991-2003) hourly precipitation produced with the same model configuration as our HR simulations [52]. The 25-km resolution for the atmosphere is likely sufficient to capture statistics of MCSs [32]. Thresholds of rain rate, area, overlapping allowed between one time-step and the next, and lifetime are used to detect organized precipitation events. Identifying MCSs therefore involved extensive sensitivity testing across parameter combinations to ensure physically consistent detection across climate states. To determine these thresholds, we randomly sampled parameter values to produce a range of statistics of detected precipitation features, which are then compared to the published MCS statistics derived from applying a well-established FLEXTRKR algorithm on radar-based observations for the same period [30]. Specifically, we calculate the root mean square error in spatial track density of MCSs across the North America between ours and published results. This allows us to determine a small set of parameters that yield low root mean square errors. Additional evaluation against ERA5 reanalysis for the same historical period shows comparable spatial patterns, further supporting the robustness of the selected parameter set (Supplementary Fig. 7). We further compared MCS rain rates and lifetimes in the NAM region for June and July between ours and published results to aid our decision (Supplementary Fig. 8). In general, our detection method shows reasonable skill at reproducing MCS spatial distribution in North America, and MCS duration, frequency, and precipitation intensity for the NAM region. Moreover, with our final parameter set, we define an MCS as a precipitating feature that contains intense rain storms (at least  $2.16 \text{ mm h}^{-1}$ ) covering a minimum area of at least 6 contiguous grid cells ( $25 \text{ by } 25 \text{ km}^2$  each) and persisting for at least 4 hours. This identification method is broadly consistent with the loose definition of MCSs [31] and was chosen to ensure systems are large, organized, and persistent, rather than isolated convection. Similar methods have been used to identify MCSs in other studies [39, 53, 54].

Our simulations and detection algorithm find that MCSs contribute up to  $\sim 60\%$  of the total precipitation in the NAM region in HR-PI (Fig. 2, d and e). Moreover, increased early summer precipitation in the HR-MP can largely be explained by more frequent MCSs (Fig. 2c) and an increase in the mean rain rate of MCSs in this region (Supplementary Fig. 9). During the early summer, the greatest number of MCSs is located along the Pacific coast in the south of Mexico, with less extending up to the northern NAM domain (Fig. 2b). MCSs become more frequent in northern Mexico in HR-MP (Fig. 2, a and c). The area with more frequent MCSs coincides with the area showing increased precipitation in HR-MP. More than  $1 \text{ mm day}^{-1}$  ( $>60\%$ ) increase in summer precipitation over the SMO in HR-MP can be explained by the MCS changes (Fig. 2f). When the same MCS detection algorithm is applied to the LR simulations, few MCSs are simulated in LR-PI and no appreciable change is detected in LR-MP (Supplementary Fig. 10), highlighting fundamental resolution-dependent differences in precipitation processes between HR and LR simulations.

## 2.3 Causes for the Pliocene MCS Changes in the NAM region and the influence of model resolution

What causes the enhanced MCS activities over SMO? To explore this question, we first calculate eddy moisture transport at sub-daily timescale to detect whether transient circulation features such as moisture surge, sea breeze or mountain breeze may have brought additional moisture to fuel the MCSs. Previous studies have shown that MCSs are strongly coupled to the eddy moisture convergence into the NAM region. As MCSs develop, the strong updrafts efficiently draw in warm, moist air from the surrounding environment, promoting airflow convergence [4, 55]. In our simulations, although HR-MP simulates an overall increase in inland eddy moisture transport from the Pacific towards the broader Mexican continent, this increase shows little spatial correspondence with where we found the most notable change in MCSs (Supplementary Fig. 11). Taken together, these results suggest that while changes in eddy moisture convergence is conducive to greater MCS activities across Mexico, it is not the primary cause for why MCS activity would be enhanced over the SMO during the mid-Pliocene. Subsequently, we focus on local processes to diagnose this cause.

To determine why enhanced MCS activity is seen over the SMO in the HR-MP, we looked at the diurnal cycle of total column convective available potential energy (CAPE) along the northern NAM transect where MCSs become more frequent (25°N-27°N, 105°W-106°W). For the NAM region, where the convective regime is relatively simple, there is a positive relationship between CAPE and precipitation: higher CAPE generally coincides with stronger convection and rainfall, as diagnosed from observations [56]. Consistent with observations, the more frequent MCSs in HR-MP found in the NAM region are associated with greater CAPE aligned with the diurnal cycle (Fig. 3, a and b). In particular, in the HR-MP, extreme CAPE events (tail end of the box plot), which are especially important for MCS formation, occur more frequently during the late afternoon and early evening compared to HR-PI (Supplementary Fig. 12), which explains the enhanced MCS activities in HR-MP. In contrast, LR simulations generally miss the tail end of high CAPE values, consistent with their inability to simulate both the mean state and the change in MCSs.

The source for the enhanced late-afternoon CAPE in the HR-MP can be traced back to the vertical structure of changes in moist static energy (MSE; Fig. 4). Following the "heat engine theory" for tropical moist convection [57], CAPE scales with the maximum transfer of MSE between the near-surface and mid-troposphere. In this framework, CAPE increases when the near-surface MSE rises independently of the mid-tropospheric MSE. Subsequently, what matters most for determining parcel buoyancy, and thus convective activities, is how much more energy the surface air has compared to the air above it (Fig. 4). In both HR-MP and HR-PI, MSE decreases vertically from the near-surface to the mid-troposphere during the late afternoon. This vertical structure indicates that near-surface air parcels are buoyant compared to their surroundings. Such conditions favor strong convection and organized storms such as MCSs. Compared to the HR-PI, HR-MP shows higher MSE through the entire tropospheric column due to the warmer and wetter climate simulated in this region (Fig. 4g). Particularly, this simulation shows a deep near-surface layer of substantial MSE increase over the SMO, which suggests the strongest accumulation of CAPE in this

area. Although similar vertical structures of MSE are present in LR simulations, the simulated near surface MSE is substantially lower in both LR-PI and LR-MP, especially over the SMO. This weaker local anomaly is consistent with the lack of high CAPE values and MCS in LR simulations in general (Fig. 3c). In sum, our simulations show that the enhanced mid-Pliocene MCS activity in the NAM region reflects a substantial increase in near-surface MSE under a convectively active climate background, both of which are only simulated at high resolution. At a low resolution, this convective background is incorrectly represented in the model as shown by the lack of high CAPE values (Fig. 3) and overall low MSE in a diurnal cycle (Fig. 4g). LR-MP also simulates notably less near-surface MSE increase compared to the HR-MP, which is insufficient to generate high CAPE values needed for MCSs development.

### 3 Discussion

#### 3.1 Climate forcing for the enhanced MCS activity in the NAM region and resolution-dependent responses

Our results suggest that the enhanced mid-Pliocene MCS activity in the NAM region arises from the locally enhanced MSE and CAPE. We argue that these changes are mostly driven by local vegetation forcing, i.e., continental greening in the NAM region [58]. In desert border regions, vegetation forcing has long been known as an important control on the precipitation regime. Charney et al. [59] first proposed that an increase in surface albedo along the desert border could decrease the absorbed shortwave radiation, resulting in reduced local MSE, and tropospheric subsidence. In contrast, a decrease in surface albedo allows for greater absorption of shortwave radiation, which tends to increase the local MSE and moist convection [59, 60]. This surface albedo-moist convection interaction potentially defines the desert boundary. The comparison between HR-MP and HR-PI is consistent with this mechanism. In HR-MP, mid-Pliocene biome distribution is prescribed based on reconstruction [46] (Supplementary Fig. 13). Nonetheless, key aspects of vegetation-climate interactions, including varying plant phenology with climate, and land-atmosphere exchange of water, energy and momentum, are dynamically simulated. Subsequently, HR-MP simulates reduced surface albedo and enhanced clear-sky net absorbed surface shortwave radiation (Fig. 5). This increase is locally amplified over the SMO. This additional net surface shortwave absorption can generate additional heat and water fluxes from the land to the atmosphere, which explains the locally enhanced MSE, and consequently, moist convection.

Why isn't this mechanism active in LR simulations? Similar mid-Pliocene and PI biome distributions are prescribed in LR-MP and LR-PI (Supplemental Fig. 13). Yet, the LR-PI featured a very different convective regime with overall low MSE and CAPE over the SMO (Fig. 3). The increases in clear-sky absorbed surface shortwave radiation (Fig. 5) and MSE are both muted in LR-MP comparing to LR-PI. Clear-sky fluxes are used here to isolate surface radiative forcing independent of cloud variability, which itself responds to changes in MCS activity. We suspect that these differences may arise from 1) better resolved grid cells with anomalously high surface shortwave radiation, which are missing from the coarser resolution, and/or 2)

more effective removal of water vapor from the tropospheric column by the enhanced convective activities. This process may act as a positive feedback on enhanced surface net shortwave absorption and MSE, i.e., greater amount of MSE may generate stronger moist convection, which, in turn, transports moisture upward and precipitates it out, leaving behind a drier and clearer tropospheric column. The associated reduction in cloud cover further allows greater surface shortwave absorption and increases MSE, which promotes renewed convection—a self-reinforcing feedback observed in modern monsoon systems [57, 59].

Although our analysis points to a key role of vegetation cover in driving MCS activity and precipitation in the NAM region, the underlying mechanisms remain uncertain. In particular, processes through which changes in absorbed shortwave radiation alters near surface MSE, especially its influences on land-atmosphere heat and water fluxes and mesoscale moisture convergence, warrant future investigation. A recent study reports that changing spatial gradients of mesoscale soil moisture may change low-level temperature contrasts and vertical wind shear, providing a key ingredient for storm growth and longevity [61]. Additional research efforts are also needed to understand the differences in how moist convection at the spatial scales between 25 km and 100 km are generated in LR and HR simulations, and corresponding differences in radiative properties. Finally, future HR experiments with varying vegetation cover would also be beneficial to understand the sensitivity of convective activities to land cover changes.

### 3.2 Implications for paleoclimate modeling

Our results highlight the importance of resolving changes in mesoscale systems in simulating mean hydroclimate during the past. High-resolution models are known to be important for capturing circulation and precipitation responses to mountain uplift, as well as their broader impact on global climate [62, 63]. High-resolution models also simulate complex changes in weather events during the past, including atmospheric rivers [64], tropical cyclones [65], and extreme precipitation [66]. Yet, the application of high resolution models in paleoclimate research remains limited, partly because proxy constraints are sparse and boundary forcings are uncertain, but such simulations are essential for understanding forced responses of the hydrological cycle as demonstrated in this study.

Furthermore, paleotempestology records (proxy reconstructions of past storms) do not exist for most of the Earth’s history, which does not allow the simulated weather events to be validated by proxy records. Nevertheless, we argue that even for the mean climate state, studying changes in past weather may help narrow the gap between proxy data and model simulations. As shown by our simulations, precipitation from transient weather systems, such as MCSs, can make a substantial contribution to the mean precipitation. High-resolution simulations therefore may show better hindcast skill than low resolution simulations because of their ability to better capture weather processes.

### 3.3 Implications for future climate projections

The influence of resolution on projected future NAM summer precipitation is qualitatively consistent with our mid-Pliocene analysis. By the end of the 21st century, a High-Resolution Model Intercomparison Project (HighResMIP) run [67], using the same model and horizontal resolution as our HR simulations, produces greater precipitation over the SMO under Representative Concentration Pathway 8.5 (HR-RCP8.5) than its LR counterpart (Supplementary Fig. 14). HighResMIP is designed to evaluate high-resolution models for present and future climate. This increase may likewise reflect better-resolved mesoscale storms, although we are not able to confirm this, as HighResMIP does not currently provide high-frequency (hourly) precipitation outputs needed to directly track MCSs. However, this interpretation is consistent with results reported by a recent study using a similar modeling framework, which find a key role of mesoscale convective systems in amplifying precipitation responses following RCP8.5, which are distinct from responses simulated by 100-km resolution models [32].

Nonetheless, unlike HR-MP, the HR-RCP8.5 does not show substantial precipitation increase over the SMO by the end of the 21st century compared to its high resolution PI run. One of the key distinctions in HR-MP is the greening of the NAM region, whereas, HR-RCP8.5 features present-day semi-arid landscape. Although land-use changes are considered in many future socioeconomic development scenarios such as RCPs [68], dynamical biome shift and range expansion are not, despite their potential importance along the desert margins [69]. Our results suggest that vegetation changes are likely important for driving future summer mesoscale storms in this region. High-resolution simulations that consider future vegetation changes may provide new insights into future hydroclimate and its related hazards in semi-arid environments.

## 4 Figures

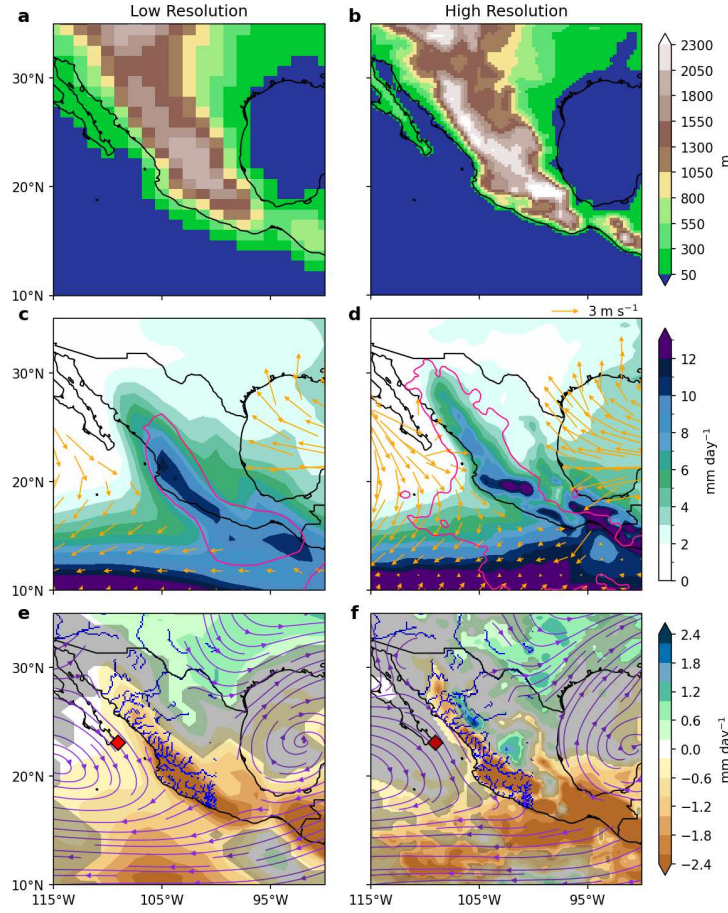


**Fig. 1 Influence of horizontal resolution in the NAM region.** (a, b) Preindustrial elevation (m) of the Sierra Madre Occidental mountain range at low-resolution (a) and high-resolution (b). (c, d) Preindustrial NAM season precipitation (June-September) (mm/day) and near-surface wind (1000 hPa; orange vectors) in low-resolution (c) and high-resolution (d). Pink contours are preindustrial monsoon extent. We define the monsoon here as summer precipitation (JJAS) exceeding 60% of the annual total. (e, f) Average early summer (June-July) precipitation difference (mm/day) between mid-Pliocene and preindustrial simulations in low-resolution (e) and high-resolution (f). Positive (negative) MP-PI values indicate wetter (drier) conditions than PI. Streamlines of wind direction differences (purple) at 850 hPa. Blue lines show the present-day river network west of 100°W. Gray shading indicates regions where MP-PI differences are not statistically significant ( $p \geq 0.05$ ). Red diamonds in (e) and (f) indicate the location of proxy site DSDP 475 [18].

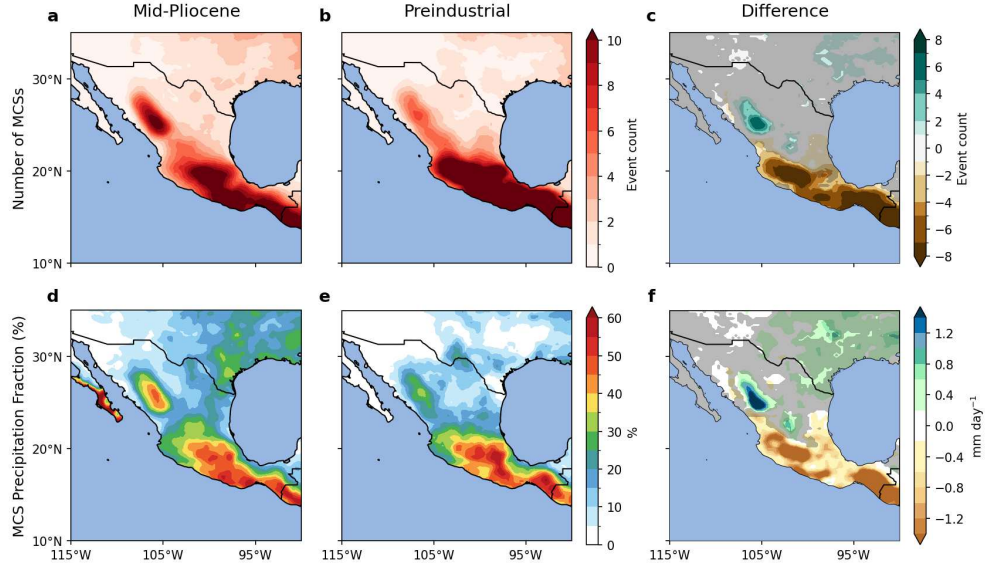


**Fig. 2 Early summer (JJ) MCS changes during mid-Pliocene.** (a-c) Average number of MCSs during JJ for the mid-Pliocene (a) and preindustrial (b) simulations and their difference (c). (d-f) MCS precipitation fraction (%) for the mid-Pliocene (d) and preindustrial (e) simulations. Difference (mm/day) between Pliocene and preindustrial in MCS precipitation (f). Gray shading in (c) and (f) indicates regions where MP-PI differences are not statistically significant ( $p \geq 0.05$ ).

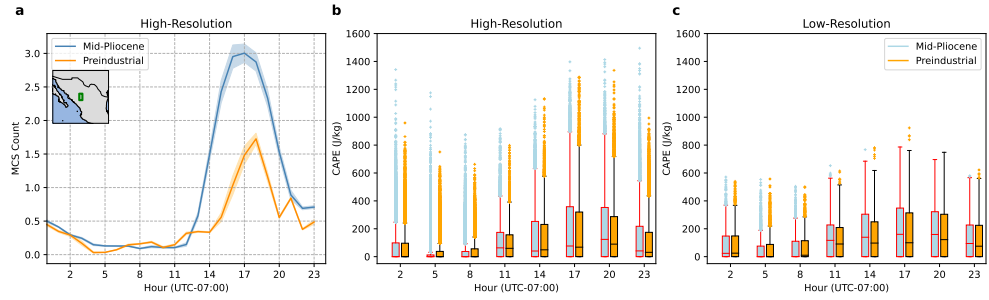


**Fig. 3 Early summer (JJ) diurnal cycle in northern NAM region (25°N-27°N, 105°W-106°W; region highlighted in inset map).** (a) Diurnal cycles of MCSs in HR-MP and HR-PI. Solid lines are mean MCS counts, and shading is +/- one standard deviation. (b) Diurnal cycle of high-resolution 3-hourly CAPE within the same region as (a). (c) Same as (b), but for low-resolution.

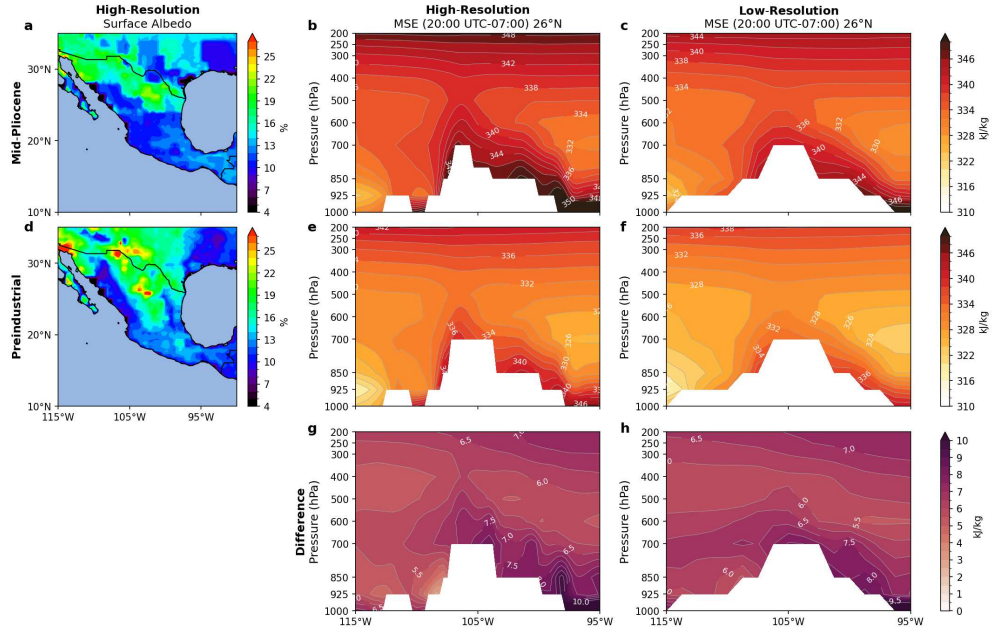


**Fig. 4 Comparison of early summer (JJ) surface albedo and moist static energy.** (a, d) Early summer (JJ) surface albedo (%) for the mid-Pliocene (a) and preindustrial (d) simulations. (b, e, g) Low-resolution early summer moist static energy (kJ/kg) during 20:00 (UTC-07:00) at 26°N for the mid-Pliocene (b), the preindustrial (e), and their difference (g). (c, f, h) Same as (b), (e), and (g), but for the high-resolution mid-Pliocene (c), preindustrial (f), and their difference (h).

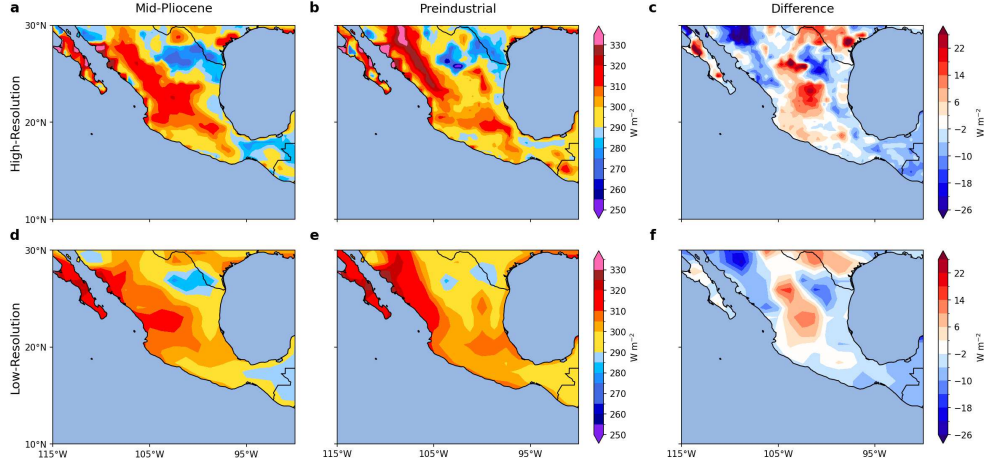


**Fig. 5 June clear sky net shortwave flux comparisons.** Clear sky net shortwave flux at the surface for (a, d) the mid-Pliocene, (b, e) the preindustrial, and (c, f) their difference in both high- (a-c) and low-resolution (d-f).

## 5 Methods

### 5.1 Experiments

Two mid-Pliocene experiments and two preindustrial (PI) control experiments were conducted using the Community Earth System Model version 1.3 (CESM1.3). They featured  $\sim 25$  km (HR) and  $\sim 100$  km (LR) horizontal grid spacing for the atmosphere and land components. Both were coupled to the same 100 km horizontal grid spacing of ocean and sea ice components. The atmospheric component used the Community Atmosphere Model (CAM) version 5 with a Spectral Element dynamical core and cubed sphere grid. On the cubed sphere grid, HR simulations used the ne120np4 grid mesh, whereas LR simulations used the ne30np4 grid mesh. The Community Ice Code version 4 was used for the sea ice component, the Parallel Ocean Program version 2 was used for the ocean component, and the Community Land Model version 4 was used for the land component. The same model and configurations have been used to study present-day and future climate [52]. LR and HR preindustrial experiments were forced with 284.7 ppm  $\text{CO}_2$  and preindustrial boundary conditions. Following PlioMIP2 protocols [26], boundary conditions for the mid-Pliocene experiments were derived from the PRISM4 reconstruction [46] and are consistent with the previous experiment using a similar model, CESM1.2 [58], except for the grid mapping. Specifically, we generated HR-MP and LR-MP topographies by applying the same PRISM4 mid-Pliocene minus PI anomaly to their respective PI baselines. This ensures that any topographic differences between HR-MP and LR-MP are solely due to resolution differences. For the SMO, PRISM4 places the peak elevation to be 229.5 meters lower than PI. Because no subgrid-scale topography variability is available for the mid-Pliocene, modern subgrid variability was mapped onto mid-Pliocene continents using

424 nearest neighbor remapping. The mid-Pliocene experiments were forced with 400 ppm  
 425 CO<sub>2</sub>. The initial conditions for ocean, sea ice, and land came from the 1200-year-long  
 426 CESM1.2 mid-Pliocene simulation [58]. By accelerating the equilibrium of ocean and  
 427 sea ice states, this initiation strategy allowed for a relatively short spin-up time (Sup-  
 428 plementary Fig. 15a). Simulation lengths varied across experiments: HR-MP was run  
 429 for 60 years, HR-PI for 100 years, LR-MP for 425 years, and LR-PI for 288 years. For  
 430 our mean state analysis, we took the last 50 years of LR and 30 years of HR simula-  
 431 tions. During these periods, both the global mean surface temperature trend and the  
 432 net top of the atmosphere radiation imbalance were negligible (Supplementary Fig.  
 433 15).

## 434 5.2 MCS Tracking

435 To track MCSs in the HR simulations, we utilized the TempestExtremes algorithm [51].  
 436 TempestExtremes uses a set of parameters to detect intense precipitation events and  
 437 determine whether they propagate. The parameters included are minimum size, min-  
 438 imum lifetime, minimum/maximum spatial overlap between timesteps, and minimum  
 439 rain rate.

440 The algorithm operates in two steps. First, it identifies “blobs” of precipitation that  
 441 meet the rain rate threshold and outputs a binary mask file. The second step connects  
 442 these blobs in time, considers the remaining parameters, and generates a file with a  
 443 unique global ID for each identified precipitation event. For our analysis, an MCS was  
 444 defined as a single entity with intense precipitation (at least 2.16 mm h<sup>-1</sup>) covering  
 445 at least 6 contiguous pixels in HR (3,750 km<sup>2</sup>) or 1 pixel in the LR simulation (10,000  
 446 km<sup>2</sup>), chosen to represent comparable minimum system scales relative to grid spacing.  
 447 It must persist for at least 4 hours, and between 10-90 % of the event’s area may  
 448 overlap from one timestep to the next. The upper overlap limit prevents stationary  
 449 or quasi-stationary rainfall features from being artificially linked into long-duration  
 450 “MCSs,” while still allowing propagation of the precipitating cores. We applied this  
 451 tracking method to all available years of 1-hourly model output: 26 years of hourly  
 452 HR-MP, 12 years of HR-PI, and 30 years of LR simulations. The LR simulations were  
 453 analyzed using the same tracking framework to assess resolution dependence, rather  
 454 than to establish a complete MCS climatology at coarse resolution. The temporal scale  
 455 of the MCS tracking in our HR simulations is limited by the amount of high-frequency  
 456 output that was saved.

457 The number of MCSs per season was calculated by counting the number of unique  
 458 MCS events in each 0.25°×0.25° pixel during June and July, then averaging across  
 459 years. Therefore, the average occurrence of MCSs per season represents the annual  
 460 count of MCSs traversing that pixel per year. The frequency and spatial arrangement  
 461 of identified MCSs in CESM historical simulations align well with findings from prior  
 462 studies on MCS climatology, e.g. [30]. Some differences from published results, such  
 463 as MCS climatologies derived by the FLEXTRKR algorithm [30] are likely indicative  
 464 of model biases. Notably, the spatial distribution of MCSs across the Great Plains  
 465 detected with our 20th century simulation shows a shift of MCS center towards the  
 466 coast compared to the distribution of MCSs detected from observational data, sug-  
 467 gesting a limitation of model skill likely in the hydrostatic balance assumption of the

dynamical core [70]. Nevertheless, in regions such as the North American Southwest, the spatial distribution of MCSs found in CESM1.3 shows acceptable agreement with previous studies (Supplementary Fig. 6; [30, 54]).

MCS precipitation fraction was calculated by extracting precipitation on days where MCSs were present for each  $0.25^\circ \times 0.25^\circ$  pixel. Total precipitation on MCS days was summed up and divided by the number of years to get the average per June–July. This calculation accounts for precipitation likely induced by MCSs but would not be detected by TempestExtremes (e.g., stratiform precipitation associated with MCS anvils). This measure was then divided by the average total precipitation for June–July to get the fraction of MCS precipitation.

### 5.3 Statistical Significance Testing

Statistical significance of mean state precipitation differences between the MP and PI simulations was assessed separately for the LR and HR model configurations. For each resolution, monthly precipitation values for June–September were compared between MP and PI using a two-sample Welch’s t-test applied independently at each grid cell. As with the mean state analysis described above, the analysis was based on 50 summers for the LR simulations and 30 summers for the HR simulations, yielding equal sample sizes within each MP–PI comparison. Welch’s t-test was used to allow for unequal variances between climate states. Regions where differences are statistically significant at the 95% confidence level ( $p < 0.05$ ) are indicated in Fig. 1e, f.

For spatial patterns of MCS occurrence and MCS-associated precipitation, statistical significance of differences between the MP and PI simulations was assessed using the non-parametric Mann–Whitney  $U$  test. The test was applied independently at each grid cell to compare the distributions of MP and PI values without assuming normality or equal variances. For MCS counts, the test was applied to inter-annual distributions of the number of unique MCS tracks intersecting each grid cell per year. For MCS precipitation, the test was applied to daily MCS-associated precipitation values aggregated across all available summers in each simulation. Regions where differences are statistically significant at the 95% confidence level ( $p < 0.05$ ) are indicated in 2c, f.

### 5.4 Eddy moisture transport

The eddy moisture transport was calculated for June and July in HR simulations following standard formulations (e.g. [71]) using sub-daily (3-hourly) model output. Prior to the calculation, a Lanczos filter was applied to remove low-frequency variability (diurnal and longer timescales), isolating transient sub-daily fluctuations relevant for MCSs; filter weights were generated using NCAR Command Language ([https://www.ncl.ucar.edu/Document/Functions/Built-in/filwgts\\_lanczos.shtml](https://www.ncl.ucar.edu/Document/Functions/Built-in/filwgts_lanczos.shtml)). Specific humidity ( $q$ ) and horizontal winds ( $u, v$ ) were decomposed into daily means and sub-daily anomalies (computed from the filtered fields), such that  $q' = q - \langle q \rangle$  and  $u' = u - \langle u \rangle$ , where angle brackets denote daily means. Eddy moisture fluxes were then calculated from the covariance of these sub-daily anomalies ( $\langle q'u' \rangle, \langle q'v' \rangle$ ) and vertically integrated through the troposphere.

509 **Supplementary information.** The supplementary materials contain Figs. S1 to  
510 S15

## 511 Declarations

- 512 • Conflict of interest  
513 The authors declare no competing interests.
- 514 • Ethics approval and consent to participate  
515 Not applicable
- 516 • Consent for publication  
517 Not applicable
- 518 • Data availability  
519 ERA5 precipitation data are available from the Copernicus Climate Data Store  
520 (<https://doi.org/10.24381/cds.adbb2d47>). IMERG precipitation data are avail-  
521 able from the NASA Goddard Earth Sciences Data and Information Ser-  
522 vices Center (<https://doi.org/10.5067/GPM/IMERG/3B-HH/06>). Mesoscale con-  
523 vective system statistics generated using FLEXTRKR were obtained from [30].  
524 The PlioMIP2 data is obtained from [72]. The CESM output generated for  
525 this study is archived on the Storrs HPC ([https://app.globus.org/file-manager?](https://app.globus.org/file-manager?origin_id=570cd647-1e13-48af-bbd3-1b201b45a14f&origin_path=%2F)  
526 [origin\\_id=570cd647-1e13-48af-bbd3-1b201b45a14f&origin\\_path=%2F](https://app.globus.org/file-manager?origin_id=570cd647-1e13-48af-bbd3-1b201b45a14f&origin_path=%2F)). The pro-  
527 cessed datasets supporting the analyses presented in this study are publicly available  
528 on Zenodo (insert link following blind review).
- 529 • Code availability  
530 MCS tracking was performed using TempestExtremes (DetectBlobs and Stitch-  
531 Blobs). TempestExtremes is open-source and available from its public repository  
532 (<https://github.com/ClimateGlobalChange/tempestextremes>). The commands and  
533 wrapper script used to generate the results, and the source code for the main results,  
534 are archived in Zenodo (insert link following blind review).

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