Climate Implications of the Messinian Salinity Crisis using the NCAR Community Atmosphere Model (CAM3.1)

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Abstract

The Messinian Salinity Crisis (MSC) occurred nearly 6 million years ago when the Mediterranean Sea (MS) was isolated from the Atlantic Ocean. High rates of evaporation and diminishing water input led to a 1500 -2500 m drop in MS level. In this study, four 20-year simulations are run using the NCAR CAM3 configured with the SOM. Two of the simulations are designed to represent specific events that occurred during the MSC. One experiment represents the complete desiccation of the MS, the period when a substantial sea level lowering occurred and anhydrites were deposited. This is known as our Lowered Land simulation. The second experiment is meant to represent the end of the MSC, known as the Lago-Mare period and is referred to as the Lowered Sea simulation.

Both of the lowered basin simulations result in anomalous wintertime warming of more than 7 K over the MS. A low-pressure anomaly develops over the basin and convergence coupled with topographic forcing yields a precipitation anomaly of more than 1 mm day^{-1} over the Alps in the Lowered Sea run. The winter storm track is strengthened near the Mediterranean Sea in both of the lowered basin runs, but is stronger in our Lago-Mare simulation. Anomalously high latent heat release due to warmer temperatures and the availability of water are believed to increase the winter storm track in the Lago-Mare simulation.

Lowering the level of the MS results in an orographically forced barotropic Rossby wave response that lead to global changes in the height field, as well as globally distributed temperature changes that are evident in both the winter and the summer.

Teleconnections extend from the Mediterranean region and result in a deepening of the Aleutian Low and a substantial cooling in the Gulf of Alaska and in the North Atlantic.

1. Introduction

a. Motivation

Nearly six million years ago (Ma) closure of the straits connecting the Mediterranean Sea (MS) with the Atlantic Ocean led to the evaporation of much of the Mediterranean, producing a subaerial depression with a depth of 1500–2500 m (Hsu et al., 1973; Barber, 1981). This spectacular event occurred in the Messinian (7 to 5.3 Ma), the last stage of the Miocene epoch (23.8 to 5.3 Ma), and is called the Messinian Salinity Crisis (MSC). Although there is a long history of utilizing numerical models to examine the influence of topography on climate (Cook and Held, 1991; Grose and Hoskins, 1978; Kutzbach et al., 1993; Manabe and Terpstra, 1974; Ramstein et al., 1997; Seager et al., 2002; Zhongshi et al., 2007), sophisticated modeling tools have not been used to shed light on the impacts the MSC may have had on global climate.

The enormous change in the level of the MS during the MSC can be expected to have brought about significant impacts on the Mediterranean climate. The basin itself must have experienced a substantial warming due to the change in elevation alone. Beyond this obvious impact, topographic changes may be expected to cause a response in the regional geopotential height field patterns that in turn imply changes in the overall regional circulation and storm tracks. Differences in storm track characteristics, such as intensity, frequency and location may result in significant changes in the hydrological cycle of the Mediterranean region (Mariotti and Struglia et al., 2001). Shifts in storm

tracks can lead to droughts or floods because storms transport copious amounts of heat, moisture and momentum (Chang, 2003). The associated changes in the local water budget may have implications in the surrounding regions by altering the availability of atmospheric moisture.

The locations of storm tracks are controlled by stationary waves. Stationary waves can be both orographically and thermally forced (Nigam and DeWeaver, 2003). Heating can affect how the flow interacts with orography, which can then affect wave trains downstream from this region (Held et al., 2002). Orography can alter the local zonal wind pattern, which can then affect the propagation of wave trains (Held et al., 2002).

Local changes in the Mediterranean region have been shown to have remote impacts in modern observations. A negative annual water budget in the Mediterranean leads to an annual mean water loss of 50-70 cm yr^{-1} (Mariotti and Struglia et al., 2001). The water budget of the MS has been positively correlated to the North Atlantic Oscillation (NAO) (Mariotti et al., 2002), a large-scale circulation pattern characterized by a seesaw in the sea level pressure between the subtropical Azores high and Icelandic low. Li et al. (2006) showed that an idealized 2 K cooling of the MS resulted in changes to both the Aleutian and Icelandic Lows (Li et al., 2006). This exemplifies the role the MS has on the environment and the teleconnectivity between the Mediterranean and remote locations.

This study examines the local response to changing sea level and removing the MS, events believed to have occurred during the MSC. In this stage of our modeling work, we focus on the atmospheric response as opposed to the coupled system because focusing on the atmosphere allows us to concentrate computational resources on

resolution and running multiple scenarios. In the future we will exam the coupled response, including the decreased salinity of the world ocean that results from the deposition of salt in the Mediterranean basin.

b. Paleohistory

The benthic $d^{18}O$ signal, a commonly used paleoclimate indicator of glacio-eustatic sea-level changes, suggest that ice formation prior to the MSC led to a decrease in global sea level by nearly 70 m (Hodell et al., 1994; Hodell et al., 2001; Clauzon et al., 1996). Changes in global sea level, orbital forcing, climate change (Hodell et al., 2001; Warny et al., 2003), and the orogenic uplift (Krijgsman et al., 1999; Duggen et al., 2003) have all been cited as closing the gateway between the MS and Atlantic Ocean during the MSC (Seidenkrantz et al., 2000; Warny et al., 2003). However, pollen data has disproved the theory that climatic variations initiated the MSC (Warny et al., 2003). Krijgsman et al. (1999; 2001) used astronomical dating methods to deduce that plate tectonics, not sea level changes, initiated the MSC by severing the connection between the MS and Atlantic Ocean at 5.96 Ma. Although, benthic foraminifera and stable isotope data show that vertical uplift as early as the latest Tortonian (approximately 11 to 7 Ma) began restricting the Rifian Corridor in Northern Africa, one of two gateways (the other being the Betic Strait in Southern Spain) that existed between the Mediterranean and the Atlantic prior to the onset of the MSC (Seidenkrantz et al., 2000; Kouwenhoven et al., 2003). Also, the benthic $d^{18}O$ signal preceded the astronomically dated initiation of the MSC (Hodell et al., 2001). Thus, while plate tectonic movements are believed to be responsible for triggering the MSC, a lowering of global sea level resulting from

glaciation may have contributed to further isolation of the MS by modulating the water level at the intervening sills (Adams et al., 1997; Aharon et al., 1993; Hodell et al., 1986).

Isolation from the Atlantic Ocean during the MSC and a negative annual water budget of the MS led to the deposition of a thick layer of evaporites (Krijgsman et al, 1999). The "Lower Evaporitic" layer includes gypsum anhydrite, halite and magnesium salts. Anhydrite and magnesium salts have been shown to only deposit in extremely hot (greater than 35°C), dry environments (Hsu, 1972; Rouchy and Caruso, 2006). The "Upper Evaporitic" and Lago-Mare (lake-sea) layer consists of gypsum and marls and marks a change in the hydrological conditions of the MS. Cooler and wetter (mostly near the mountainous regions) conditions are believed to have existed during the Lago-Mare event (Rouchy and Caruso, 2006).

The Messinian was marked by 18 glacial-interglacial cycles (Hodell et al., 2001) and three peaks of glaciation occurred just after the initiation of the MSC (Vidal et al., 2002). Ice rafted detritus (IRD) data from Site 646 has indicated sea-ice cover from the Labrador Sea to south of Greenland and glaciation in the high altitudes of Iceland in the latest Miocene (Jansen and Raymo, 1996; Mudie et al., 1983; Vidal et al., 2002). IRD dating from the Latest Miocene was also discovered in the Yakataga Formation, a 7 km thick sediment core from Ocean Drilling Project (ODP) Site 887 in the Gulf of Alaska (Zellers et al., 2007). This represents localized glaciation in both the North Atlantic and North Pacific during the period of the MSC. Sun and Liu (2006) suggested that a 6‰ reduction in global oceanic salinity during the MSC (Hsu et al., 1977) may have resulted in greater sea ice formation due to the higher freezing point of the lower salinity water, an effect we hope to investigate with fully coupled simulations in future work.

2. Methodology

a. Model

In this study, we used the NCAR CAM version 3 (Collins et al., 2004), a well-documented atmospheric general circulation model, configured with a Slab Ocean Model (SOM). Although SSTs are able to respond to changes resulting from a modified MS and atmospheric changes, the SOM does not simulate the full ocean circulation. For the local changes in the Mediterranean climate investigated here, ocean circulation feedbacks are unlikely to have had a first order effect on the change due to the MSC. However, ocean-atmosphere feedbacks are more important for the remote effects seen throughout the Northern Hemisphere.

The atmospheric model uses a terrain following hybrid coordinate with a spectral Eulerian dynamical core and is comprised of 26 vertical levels. In this study, it is coupled to a mixed layer SOM that is combined with a thermodynamic sea ice component. The SOM configured CAM3 allows for the inclusion of a prognostic mixed layer temperature rather than prescribed SST. Horizontal oceanic heat transport is represented by compensating heat sources and sinks known as the Q flux, which are derived by running a fixed-SST model to equilibrium, and setting the Q flux equal to the local heating or cooling in the ocean mixed layer that compensates for loss or gain of energy to the atmosphere via the modeled aerodynamic fluxes. For these experiments CAM3 is run at T42 resolution (an equivalent grid spacing of roughly 2.8^{o} x 2.8^{o}) (Fig. 1). CAM3 is coupled to the Community Land Model (CLM3) (Oleson et al., 2004), which incorporates biogeophysics, hydrological cycle, biogeochemistry and dynamic

vegetation. The vegetation is divided into plant functional types (PFT) that are characterized by its structure. The vegetation structure, including leaf and stem area index as well as canopy height, is input to each grid cell for each PFT. Greater detail of this model is provided in Collins et al. (2004) and Oleson et al. (2004).

b. Experimental design

Four experiments were performed using the NCAR CAM3 to simulate the series of events that occurred during the MSC and isolating the effect of each separate component on the atmosphere (Table 1). Heat flux is normally transported into the Mediterranean Sea from the Atlantic Ocean. During the MSC, the Mediterranean Sea was isolated from the Atlantic and therefore it is assumed that heat was not transported into this basin. For each experiment, the horizontal ocean heat transport (Q flux) into the Mediterranean Sea is shut down. This represents the isolation of the MS from the Atlantic Ocean during the MSC. The SST dataset in the SOM requires two additional fields specifying mixed layer depths and Q-fluxes. Q flux information was gathered from a run in which we converted the MS to land and ran the CAM3 with a data ocean model for 10 years with climatological blended HadISST and Reynolds SST datasets. Therefore, the surface flux balance information from the data ocean run were then used to compute the new Q fluxes needed to run the SOM. Eliminating horizontal heat transport into this region through the strait was also done to ensure that we isolate the impact of changing the MS level.

Trace gas concentrations, including atmospheric CO_2, are set to present day levels based on alkenone-based estimates of pCO_2 (Pagani et al., 1999) that indicate pCO_2

values were above pre-industrial values during the late Messinian. The default CO_2 concentration in CAM is 350 ppm. Orbital parameters and the solar constant were also kept at present-day values. Using present-day levels of CO_2 and insolation allows us to isolate the effects of changing topography and land characteristics. Since only minor plate tectonic movements have existed since the Tortonian (Ruddiman and Kutzbach, 1989; Prell and Kutzbach, 1992; Ramstein et al., 1977) we followed the Francois et al. (2006) and Stepphun et al. (2006) studies and assumed a modern day land-sea distribution for our Messinian runs. Modern day vegetation is also used in this study.

In our control upper sea (US) run, we simulate the impact of isolating the MS from the Atlantic Ocean. The new horizontal heat transport information is implemented, but the sea level and surface characteristics of the MS are unchanged. Changing the horizontal heat transport in the ocean, Q flux, is assumed to have several important implications for global climate in the control run (US). Normally, the Mediterranean Sea is a substantial source of heat for the overlying atmosphere. When the Mediterranean Sea is disconnected from the Atlantic Ocean, the missing heat flux is redistributed globally and can lead to deviations from a true control CAM3 SOM run. A sensitivity test is also conducted to investigate the effect of changing land surface characteristics and not topography over the MS. In this experiment, known as the upper land (UL) run, we convert the MS to land but the surface remains at present day sea level. Changing the land surfaces causes changes in the fluxes of heat, momentum and water to the overlying atmosphere. This is incorporated in the Community Land Model version 3 (CLM3) that is run with the stand-alone CAM (Oleson et al., 2004). At T42 resolution only three grid points over the MS are considered to be completely ocean and the other 31 Mediterranean

grid boxes are partial land-ocean points (Fig. 1). In order to remove the Mediterranean Sea these ocean and partial ocean-land boxes are converted to land. In order to account for the change from ocean to land, soil color, texture and plant characteristics were added. These values were taken from a grid box over the Saharan Desert at 8°E and 18°N. We assumed that the land characteristics of a desiccated basin would be similar to a point representing a desert.

In the lowered sea (LS) experiment, the MS level is reduced up to a maximum of 1500 m, a median estimate of sea level fall during the MSC. The elevation was lowered as a linear function of land fraction. Complete ocean points were lowered the full 1500 m whereas fractional land/ocean points were lowered less than 1500 m. Much of the surface of the basin is between 700 m to 1400 m below sea level with only three grid points in the center of the MS that reach a maximum depth of 1500 m. In this experiment, the Mediterranean sea level is lowered, but water is still retained in the basin. This was done to examine the effect of sea level fall on the climate and is representative of Lago-Mare conditions that were present at the end of the MSC.

Next, the surface of the MS is converted to land and is lowered in the same manner as the LS simulation. We refer to this experiment as the lowered land (LL) experiment and the basin is fully desiccated. This represents the period when the "Lower Evaporite" layer was deposited and extremely dry conditions prevailed. Surface characteristics are the same as the land sensitivity (UL) run. This experiment examines the effect of loss of water and sea level fall in the MS.

All four experiments were run for 70 years. Climatological annual and monthly means are calculated over the last 30 years of the model runs. The climatological monthly

means were then used to calculate monthly climatological standard deviations. A student's t-test was performed to determine where the experiments were statistically different from the control run and values that were statistically different at the 0.05 or less critical level of significance are stippled on the difference maps.

3. Results

Lowering the surface topography of the Mediterranean Sea results in a 160-hPa annual mean surface pressure anomaly (Fig. 2) over the Mediterranean basin relative to our control. This leads to a lowering of the height field that is evident throughout the troposphere above the basin. The height field at 500 hPa is lowered 1000 m and at 300 hPa it is lowered 800 m. The atmospheric response to this substantial elevation change over the Mediterranean Sea is dependent on the surface characteristics as well as the season. In the following discussion we discuss anomalies between the three experiments (LS, LL, and UL) and our control run (US).

A true CAM3 SOM control run is not used to compare to our results because our new heat flux implementation leads to deviations from a real control run (discussed further in Section 3c). It is assumed that the missing heat transport into the Atlantic is redistributed globally throughout the oceans; however, fully coupling CAM to an ocean general circulation model is believed to lead to more accurate results. New runs using CCSM3 are being prepared.

a. Climate response during the winter

i. Temperature response

Lowering the sea level of the MS (LS case) leads to a distinct warming of the Mediterranean basin (Fig. 3). Assuming an adiabatic lapse rate from sea level down to the lowered Mediterranean predicts a warming of 14 K. We expect the warming to be somewhat less than this, since the terrestrial lapse rate is strongly governed by the surface energy budget, which is not strongly height-dependent (Meyer, 1992). Figure 3a shows the December-January-February (DJF) wintertime warming over the basin. We would expect the LS basin to be warmer than the LL basin because of the higher heat capacity of water that should allow the basin to retain the heat that was gained during the summer. The DJF warming above the lowered Mediterranean is independent of the surface characteristics. Both the Lago-Mare and complete desiccation simulations lead to a similar DJF warming of more than 7 K. Thus, it is the heat capacity of the overlying atmosphere that prevents temperatures from rising much higher during the day in the Lago-Mare simulation.

Meridional (north-south) vertical cross sections of seasonal temperature anomalies show that anomalously warm air extends into the upper troposphere over a lowered basin. This warm bubble of air is similar in both of the lowered surface runs and extends to nearly the same altitude of 200 hPa. Changes in advection by the mean winds carries warm air into northeastern Africa. Northeastern Africa is 0.5 K warmer than our control (Fig. 4).

ii. Circulation and atmospheric water balance

The large warming that occurs when the elevation of the Mediterranean Sea is lowered changes the local wind pattern (Fig. 5a). The wind field shows anomalous cyclonic (counterclockwise) rotation at the surface, which is representative of low

atmospheric pressure. In contrast to the simulations where the height field is unchanged and the winds over the MS are primarily zonal (flowing from east to west), the Lago-Mare and complete desiccation simulations produce a strong southeasterly wind over the central and eastern MS. The vertical velocity field shows subsidence (sinking) in regions where the wind is entering the basin. This air is being pushed down the slope at the northwestern edge of the MS and is pushed up the slope in the northeastern edge of the MS. Vertical ascent is evident in regions where air is exiting the basin. Changes in the wind pattern drive changes in the hydrological cycle (Fig. 6a). A significant increase in precipitation occurs along the northeastern margin of the MS. Precipitation is enhanced in this area because the southeasterly wind pushes air up the slope of the northern edge of the basin. The winter climate of the Lago-Mare scenario results in a topographically forced precipitation anomaly of 1 mm day^{-1}. This precipitation anomaly extends from Italy to Bulgaria. The precipitation anomaly for the complete desiccation scenario is located in the same region but is only half as strong.

A moderate but significant increase in evaporation along the southern edge of the sea occurs in the Lago-Mare scenario. E-P varies from positive values in the south to negative values in the north of the MS (Fig. 7a). The moisture budget over the Mediterranean watershed depends on the direction the wind is blowing. If the wind is being directed over the MS then it will entrain moist air and transport this air downstream where it can condense out. In our modeling study, when sea level is lowered, the highly saturated Mediterranean air is transported towards the northwest by the mean winds. This is due to a westward shift of the surface low in response to the strong surface heating (Fig. 2). Southerly flow would transport warmer, more humid air into southern

Europe. Since the absolute humidity, or vapor density, of warmer air is higher than for cooler air the warmer air is able to hold more moisture. This moisture is then released when the air is orographically lifted at the northern margin of the basin. The precipitation anomaly in the complete desiccation scenario is smaller than the Lago-Mare due to the lack of water. When the MS is removed and replaced with land (LL and UL runs), precipitation is greatly reduced from the eastern basin into the Middle East. The water budget (E-P) when the sea is removed shows positive values in the Middle East and negative values over the sea.

The impact of the desiccation of the MS on the annual atmospheric water balance integrated over the entire Mediterranean drainage basin (0-40°E, 20-50°N) is shown in Table 2. This region covers the Nile River Delta, the Alps and the Black Sea. Our simulation of the Lago-Mare period results in an annual E-P value of 3.1 mm day^{-1}, a value that is similar in magnitude to the control climate. It is important to note that the Dead Sea doesn't evaporate completely because its increased salinity reduces the equilibrium vapor pressure of water over it (Raoult's Law), reducing evaporation. A very salty Lago Mare might have somewhat lower E-P than our results indicate. Rouchy and Caruso (2006) argue that climate change led to the transition from Lower to Upper evaporite deposition and involved a moistening of the mountainous regions surrounding the Mediterranean basin. In our Lago-Mare simulation the precipitation is shifted to a new location although the total water budget of the Mediterranean watershed remains unchanged. However, when the MS is completely desiccated (LL run), the integrated Mediterranean watershed E-P reduces to 0.33 mm day^{-1}.

This shows that the complete drying of the Mediterranean basin causes a sufficient reduction in evaporation to the point where the E-P budget of the basin is brought very close to balance. A small externally forced increase in precipitation could easily tip the integrated E-P to a positive range, partially refilling the basin, and initiating Lago-Mare conditions. If salinity is not a substantial barrier to evaporation, the basin is unlikely to refill completely, since a substantially filled basin (as in the Lago-Mare simulation) substantially increases evaporation to the nearly control conditions, preventing complete refilling until some larger climatic event occurs, or until a tectonic event reopens the connection with the Mediterranean.

ii. Rossby wave response

Anomalously low pressure inside the deep basin should lead to an increase in storminess, if other factors are held constant. The kinetic energy due to transient eddies, or Eddy Kinetic Energy (EKE) at 250 hPa, is a good diagnostic for storm tracks. EKE is defined as

$$EKE = \frac{1}{2}(u'^2 - v'^2)$$

where $u' = u - \langle u \rangle$ represents a deviation from a time average and $\langle u \rangle$ represents a time average of u. EKE at 250 hPa in our Lago-Mare scenario shows a significant increase near the Mediterranean region during the winter (Fig. 8). The presence of both warm temperatures and abundant water in our Lago-Mare simulation results in higher latent heat flux over the MS (Fig. 9a). Increased latent heat flux suggests that there is more energy readily available in the atmosphere to generate and sustain weather disturbances. In the complete desiccation scenario the increase in storminess near the Mediterranean is not as great. We can deduce that the presence of anomalously high latent heat release

contributes much more to the storm track near the MS than the presence of anomalously low pressure because the deviations in surface pressure are consistent between both simulations. Variations in the storm track can lead to transient eddy feedbacks, which can alter the Hadley circulation. This in turn impacts the storm track (Liu et al., 2007), both of which are key components in the atmospheric circulation.

Large changes in the boundary conditions of the MS produce an orographically forced barotropic Rossby wave response that results in changes in both the height field and temperature field throughout the Northern Hemisphere (Figs. 10-11a). Figures 10-11a display the geopotential height response for all three experiments at 850 hPa and 300 hPa, respectively. The development of a wave train that propagates throughout the Northern Hemisphere is apparent in the 300 hPa maps of geopotential height anomalies. Monthly averages of geopotential height show maximum variance near the semi-permanent lows and explain a substantial portion of the wintertime atmospheric circulation. A difference in the average geopotential height field indicates variability in the circulation and therefore changes in the paths that storms tend to follow.

During the winter, both the Lago-Mare and complete desiccation simulations result in a barotropic response over the Mediterranean Region with vertically stacked low height anomalies from the surface to the upper troposphere. Lowered heights extend westward across the North Atlantic into North America. Higher heights occur downstream from the MS over Russia and Eastern Europe. At 300 hPa the height field is lowered by about 80 m in the west and raised by about 80 m in the east. A stronger Rossby wave response (evident in the 300 hPa anomalies) occurs in our Lago-Mare simulation and is a consequence of a stronger storm track in the North Atlantic.

A deepening of the Aleutian Low is seen in both the Lago Mare and complete desiccation scenarios. These changes are evident in the January geopotential height field, which shows a strong decline in the height field over the Gulf of Alaska (Fig. 11a). A corresponding 2 hPa drop in sea level pressure occurs just south of the Aleutian Islands. The Aleutian and Icelandic lows are both semipermanent lows through which synoptic scale low-pressure systems frequently pass and intensify (Zhu et al., 2007). Both the Aleutian Low and Icelandic Low are the most intense during the winter when the baroclinicity, or equator-to-pole temperature gradient, is largest. These centers have significant effects on the atmospheric circulation in the Northern Hemisphere by marking the front between subpolar and subtropical air.

A strengthening of the EKE at 250 hPa occurs in the eastern North Pacific storm track region indicating increased storminess (Fig. 8). Anomalously lower pressure in the Gulf of Alaska, resulting from increased storms, results in cyclonic rotation of the surface wind for both the Lago Mare and complete desiccation simulations (Fig. 12). In the extratropical oceans, cyclonic rotation leads to divergence at the surface and subsequent upwelling of cold water from the deep ocean. This results in a statistically significant cooling of more than 0.5 K over the north Pacific (Fig. 4a). This result agrees nicely with IRD data from ODP Leg 145, which gives evidence of localized glaciation in the Gulf of Alaska during the latest Miocene (Zellers and Gary, 2007).

b. Climate Response during the summer

i. Temperature response

The June-July-August (JJA) air temperature above the LL basin is several degrees warmer than the LS basin because of the smaller heat capacity of the LL case (Fig. 3b). In both the Lago-Mare and complete desiccation scenarios a warm air bubble extends into the stratosphere. At 100 hPa there is a temperature anomaly of 1 K. When the Mediterranean Sea is removed (LL and UL runs), the anomalously warm air extends southward into northern Africa at low levels. Cooling occurs in the mid-Troposphere near 500 hPa in both the Lago-Mare and complete desiccation simulations. This cooling to the south of the MS leads to a strong meridional gradient in temperature in the mid-Troposphere.

ii. Circulation and atmospheric water balance

Cyclonic rotation develops in the northwest quadrant of the basin and anticyclonic rotation develops in the southeast quadrant of the basin (Fig. 5b). This creates an intense southwesterly wind that blows across the lowered MS. Southwesterly winds carry hot air into Eastern Europe and Russia (Fig. 4b). Similar to the DJF circulation in the basin, subsidence occurs in the northwest as air enters the basin and vertical ascent occurs in the southeast as air exits the basin. In the Lago-Mare scenario, there is decreased summer precipitation, less than 1 mm day^{-1}, in the west where there is increased subsidence (Fig. 6b). Typically, a Mediterranean climate consists of warm, dry summers, however, due to topographic lifting in our Lago-Mare scenario JJA precipitation is enhanced by more than 0.5 mm day^{-1} near Greece. The lack of moisture inhibits precipitation in the complete desiccation scenario even though there is vertical ascent. The atmospheric water budget (E-P) shows much less evaporation during the summer when water is removed from the Mediterranean (Fig. 7b).

ii. Rossby wave response

The downstream JJA response is weaker than the DJF response, but shows the same characteristics (Fig. 10-11b). Lower heights and sea level pressure occurs near Alaska. This feature occurs throughout the year when the elevation of the Mediterranean Sea is lowered. Low heights also extend across the North Atlantic reaching the North American continent. This leads to cooler SSTs in the North Atlantic as well as the Gulf of Alaska (Fig. 4b).

The height field anomalies at 300 hPa are much stronger than at 850 hPa. Low level warming in northeastern Africa leads to increased surface heights at 850 hPa. This feature is not seen at 300 hPa. This dipole pattern in the geopotential height field at the surface (low heights over the Mediterranean Sea and high heights to the southeast over Africa) may provide enough forcing to the atmosphere to create a large-scale response during the summer when the storm track is weaker.

c. The impact of changing Q-flux

A comparison of our simulations (US, UL, LS, and LL) to a Eulerian CAM3 SOM control simulation (CNTL) at the same resolution (T42) shows significant deviations in the global climate that is not confined to the perturbation region. Our calculated oceanic heat flux is the only consistent change in our three similations, therefore similar changes in all three difference plots (US-CNTL, LS-CNTL, LL-CNTL,UL-CNTL) can be related to the difference in the oceanic heat flux used in our simulations compared to a control run. Eliminating oceanic heat flux into the Mediterranean Sea results in cooler tropical oceans, especially around the cold tongue

regions in the Pacific and Atlantic, and warmer high latitude oceans (Fig. 13). The zonal average of the difference in surface temperature shows that our new heat flux scheme has a substantial cooling effect (near -1 K) on the latitudes that enclose the Mediterranean Sea and a strong warming (near -1 K) in the Northern Hemisphere high latitudes (not shown). The zonal average temperature shows cooling along 40°N in spite of the large warming over the basin in the LS run. The zonal average of the difference in sea level pressure between the control and our three experiments shows that the LL and LS runs results in lower pressure throughout the tropics and midlatitudes.

Examining the variations between the difference plots ((US-CNTL and UL-CNTL) versus (LL-CNTL and LS-CNTL)) allow us to determine the consequences of the new surface topography that is implemented over the Mediterranean Sea. This shows that the sea level pressure changes in the Northern Hemisphere result from a Barotropic Rossby wave response that is generated because of the large surface depression over the Mediterranean Sea (Fig. 14). This Rossby wave response is responsible for the deepened Aleutian Low and cooler Gulf of Alaska, most notably in the summer. During the winter, the US simulation has higher pressure over Alaska but the response of changing the surface topography (LS and LL) opposes this effect.

4. Discussion

In our Lago-Mare and complete desiccation simulations, we concluded that changes in the zonal wind pattern in a partially desiccated basin led to a large positive anomaly in precipitation that extends from the Alps eastward into Bulgaria during boreal winter. The precipitation anomaly over the Alps is greater than 1 mm day^{-1}. This

anomaly is substantial when compared to climatological area averaged precipitation, which varies between 1.4-4.8 mm day^{-1} during the winter in this region (Mariotti and Struglia 2001). Willett et al. (2006) suggested that wetter conditions during the Lago-Mare period led to the abrupt cessation of outward growth in the Alps. Our results suggest that a switch in the zonal winds due to the lowered elevation of the MS during the Messinian coupled with the availability of water could allow air parcels to entrain water as they move across the lowered basin. This water can later be precipitated out as air is forced up the sloping surface in the northeastern edge of the Mediterranean Sea. The transport of water from the MS to the Alps during the Lago-Mare period may have led to increased weathering. Rouchy and Caruso (2006) state that the transition to the Lago-Mare period occurred when a wetter climate existed near the mountainous regions along the coast of the MS. In our Lago-Mare simulation, a significant 1 mm day^{-1} increase in precipitation occurs along the northern edge of the MS where air is pushed up the slope of the lowered basin. This result tends to support these conclusions. A strong gradient in the E-P field occurs over the lowered MS in our Lago-Mare simulation. A strong negative anomaly of E-P occurs along the northern edge with a maximum centered over the Alps and positive values along the southern margin of the basin. The positive E-P values along the southern margin of the MS results from an increase in evaporation. This result opposes Griffin's (2002) claim that saturated MS air would be transported into Northern Africa thereby strengthening the North African monsoon during the MSC.

Our results show cooling of high latitude ocean regions. There is support for this in the geological record. Strong cooling over the North Pacific, specifically off the coast of the Aleutian Islands, occurs throughout the year in the Lago-Mare and complete

desiccation scenarios. Lowering the sea level of the MS results in a cooling of more than 2 K in the North Pacific. This is consistent with IRD evidence from the Yakataga Formation in the far North Pacific Ocean and from ODP Leg 145 in the Gulf of Alaska, which recorded localized glaciation during the latest Miocene (Zellers and Gary, 2007). Strong and statistically significant cooling of the North Atlantic SSTs occurs in the Lago-Mare and complete desiccation scenarios. Utilizing a coupled model may result in important feedbacks in the thermohaline circulation as it is expected that the salinity of the Atlantic Ocean will be reduced due to the absence of Mediterranean outflow water. Lower salinity and cooler waters in the North Atlantic would reduce the overturning circulation and lead to important consequences to global climate.

5. Conclusions

Lowering the MS level led to a 160-hPa decrease in the annual surface pressure relative to the control run. Strong adiabatic warming occurred as air rushed into the deep basin. During the summer, our simulation of the complete desiccation of the MS resulted in the largest warming. In JJA, the surface temperature over the basin warmed by more than 12 K. In DJF, the Lago-Mare and complete desiccation simulations showed similar warming of 7 K. Anomalous warming in the Lago-Mare simulation resulted in changes to the regional circulation and combined with orographic lifting resulted in a 1 mm day^{-1} increase in precipitation near the Alps. The combination of both warm temperatures and the abundance of water in the winter climate of the Lago-Mare simulation led to a strong increase in the latent heat flux. Increased latent heat flux resulted in an intensified Mediterranean storm track because more energy was available for storms. This led to

transient eddy feedbacks, which have been shown to modify the Rossby wave response (Watanabe and Kimoto, 1999).

We showed that lowering the sea level of the MS did not change the annual water balance of the Mediterranean drainage basin, but did result in the redistribution of precipitation around the Mediterranean drainage basin. The complete desiccation of the MS, however, resulted in a significant reduction in the integrated Mediterranean watershed E-P. Therefore, the extremely arid climate that favored the deposition of the Lower Evaporite layer must have been externally driven.

A schematic illustration of the global climate effects of the MSC is shown in Figure 15. The simulated climate implications of the MSC are not confined to the Mediterranean region, but instead propagate globally through variations in the storm track. A global barotropic downstream response occurs throughout the Northern Hemisphere during both seasons when the elevation of the MS is lowered (LS and LL) but is strongest in the winter when the storm track is intensified. This Rossby wave response propagates downstream and initiates global changes in the height field. These changes lead to strong temperature anomalies in both winter and summer that are especially large over the North Atlantic and North Pacific. In the Lago Mare and complete desiccation simulations a remarkable deepening of the Aleutian Low occurred.

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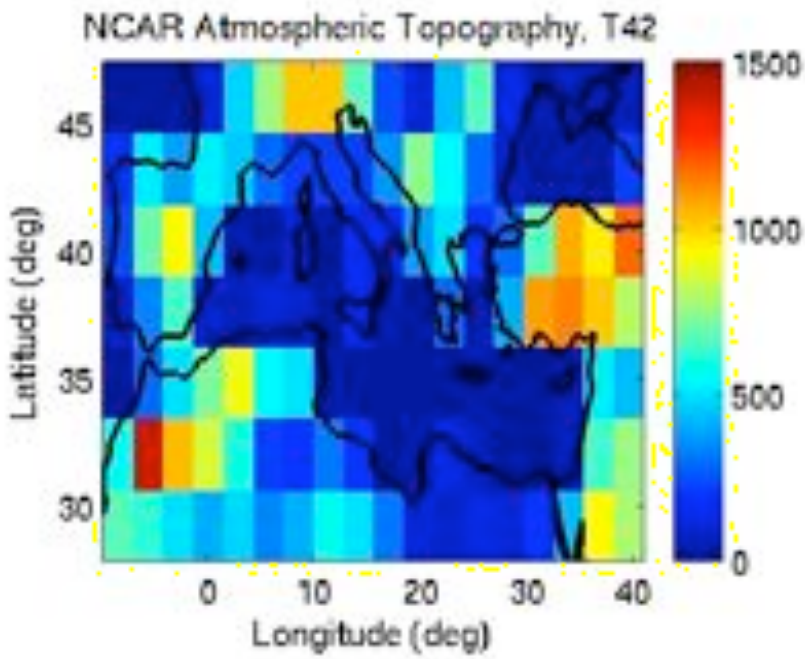


Figure 1. Topography over the Mediterranean region as represented by T42 resolution in the NCAR CAM 3.1.

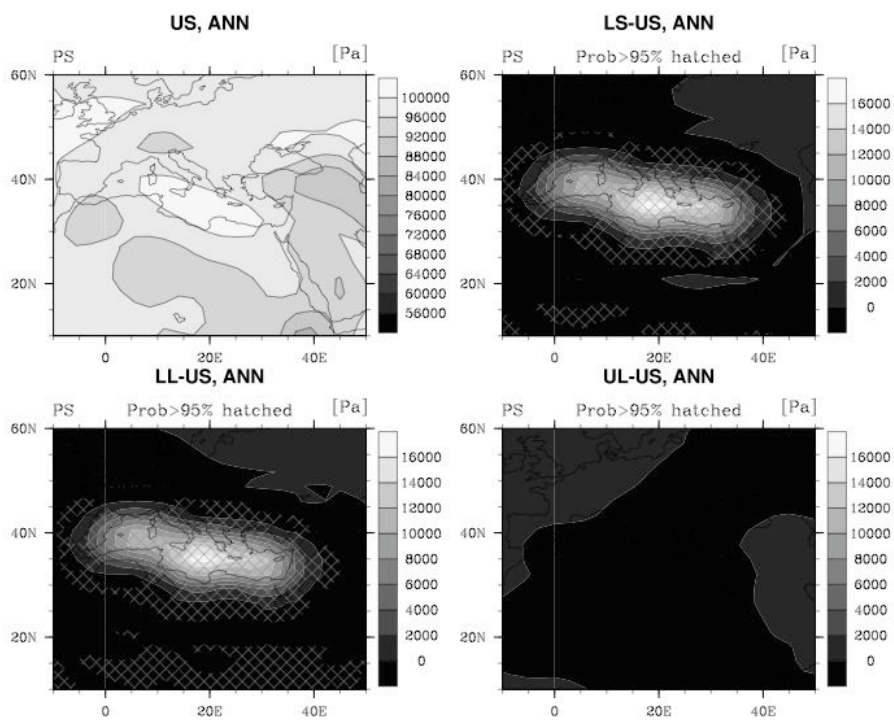
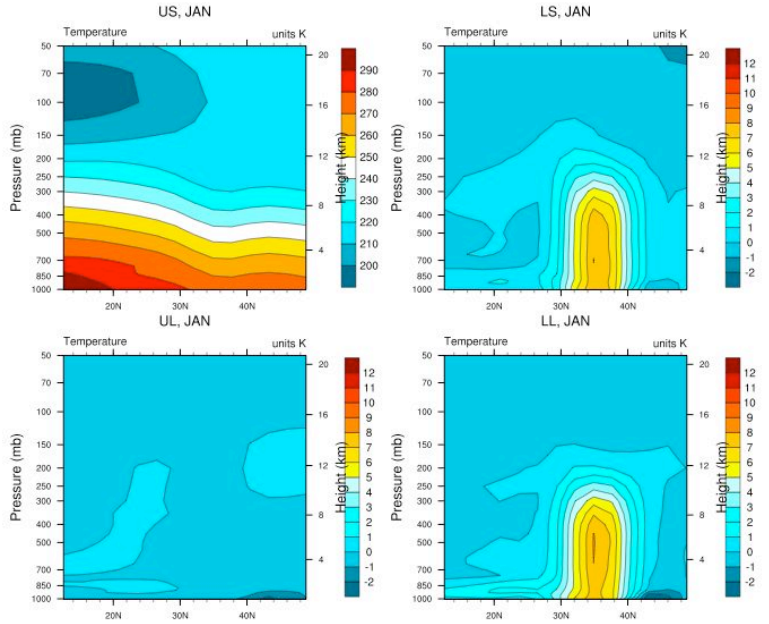
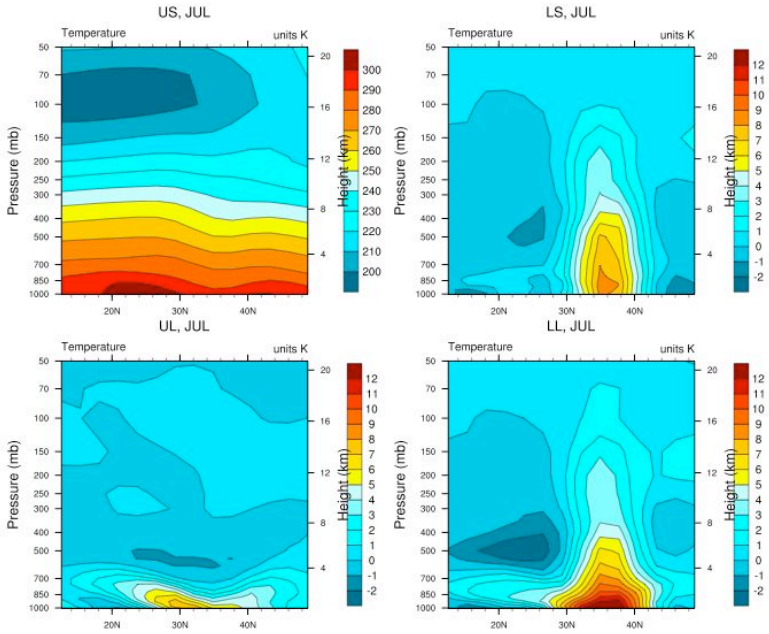


Figure 2. Annually averaged surface pressure (top left panel) and surface pressure anomalies (next three panels) in Pascal's. The top left panel is the control (US) case. The anomaly plots are stippled where differences are significant at the 95% confidence level.



A.



B.

Figure 3. Meridional (north-south) vertical cross section of temperature for A) January and B) July. The top left panels of A and B show the control climate (US) and the next three panels show the differences between each simulation (LS, UL, and LL) and our control (US).

A. **B.**

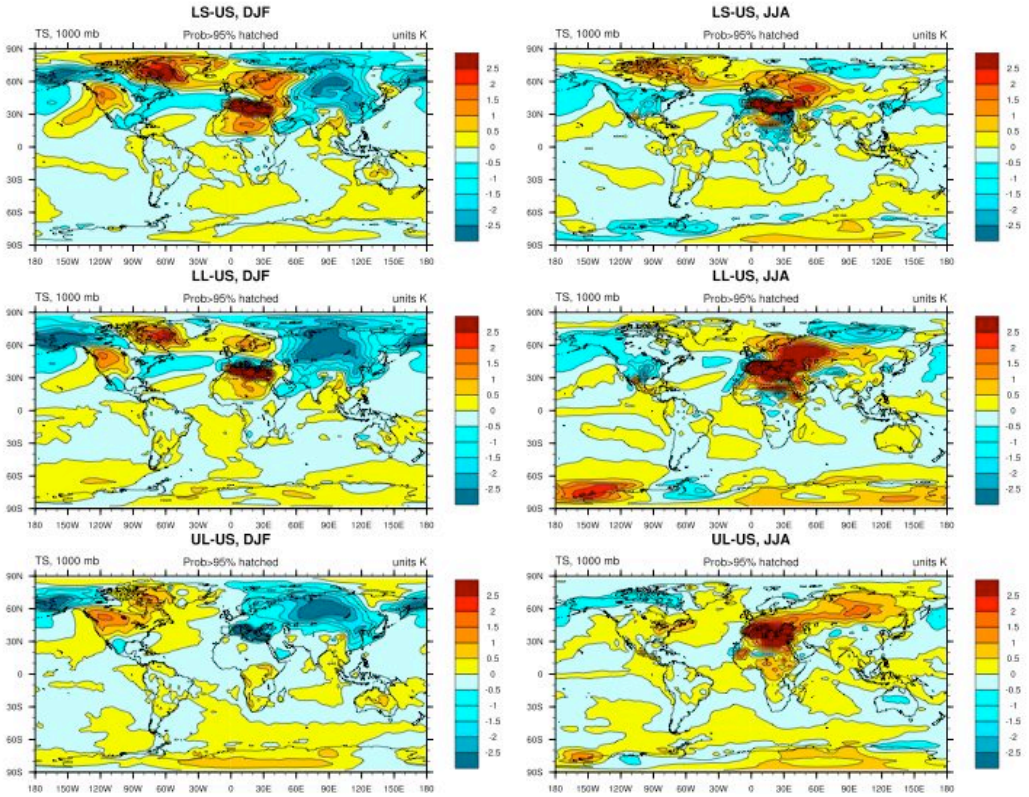
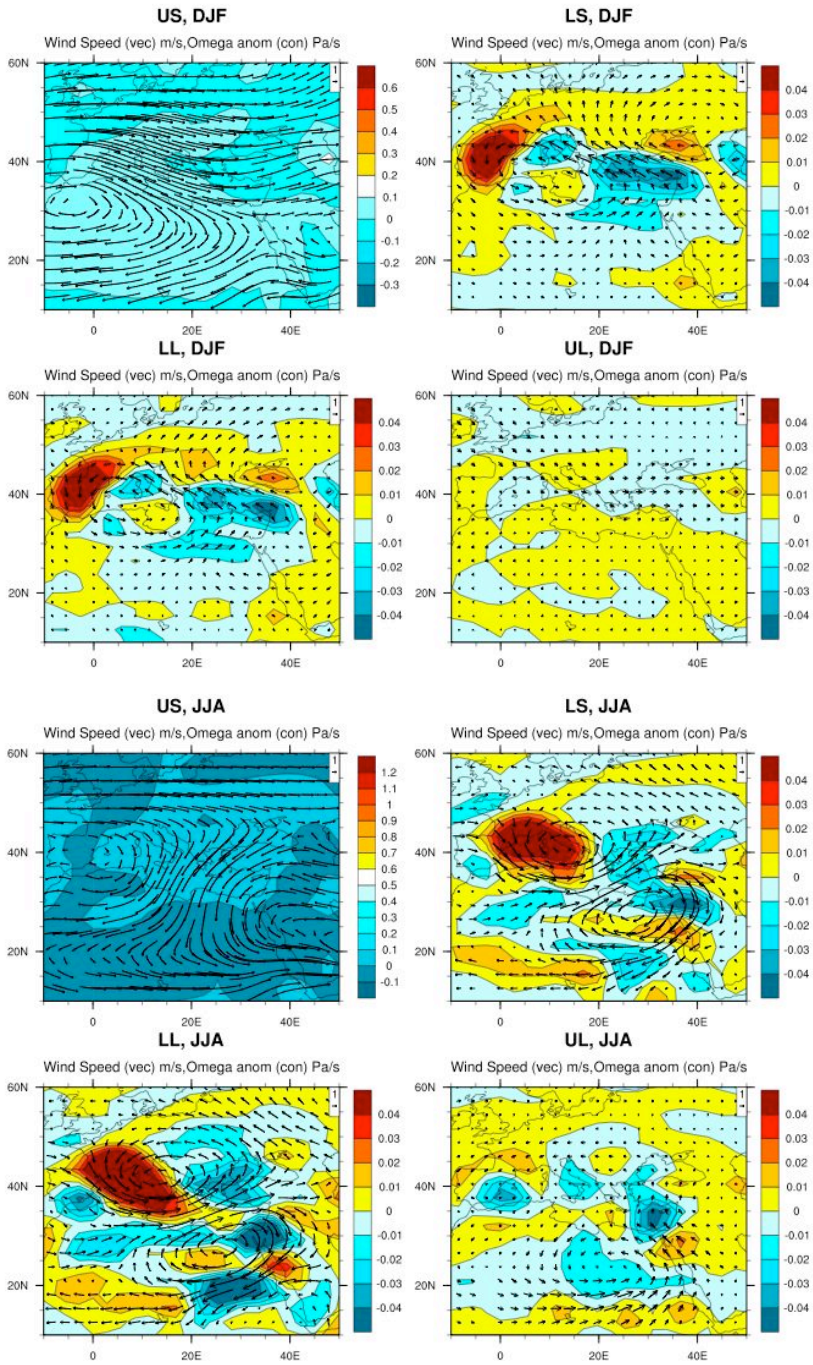


Figure 4. Surface temperature (TS) anomalies in Kelvin averaged over the last 10 years of model run. Plot A shows the climatological DJF difference plots and plot B shows the climatological JJA difference plots.



A.

B.

Figure 5. Vertical velocity (shaded), in Pa s^{-1}, and wind (vectors) over the Mediterranean watershed for a) January and b) July. The top left panels of A and B show the control climate (US) and the next three panels show the differences between each simulation (LS, UL, and LL) and our control (US).

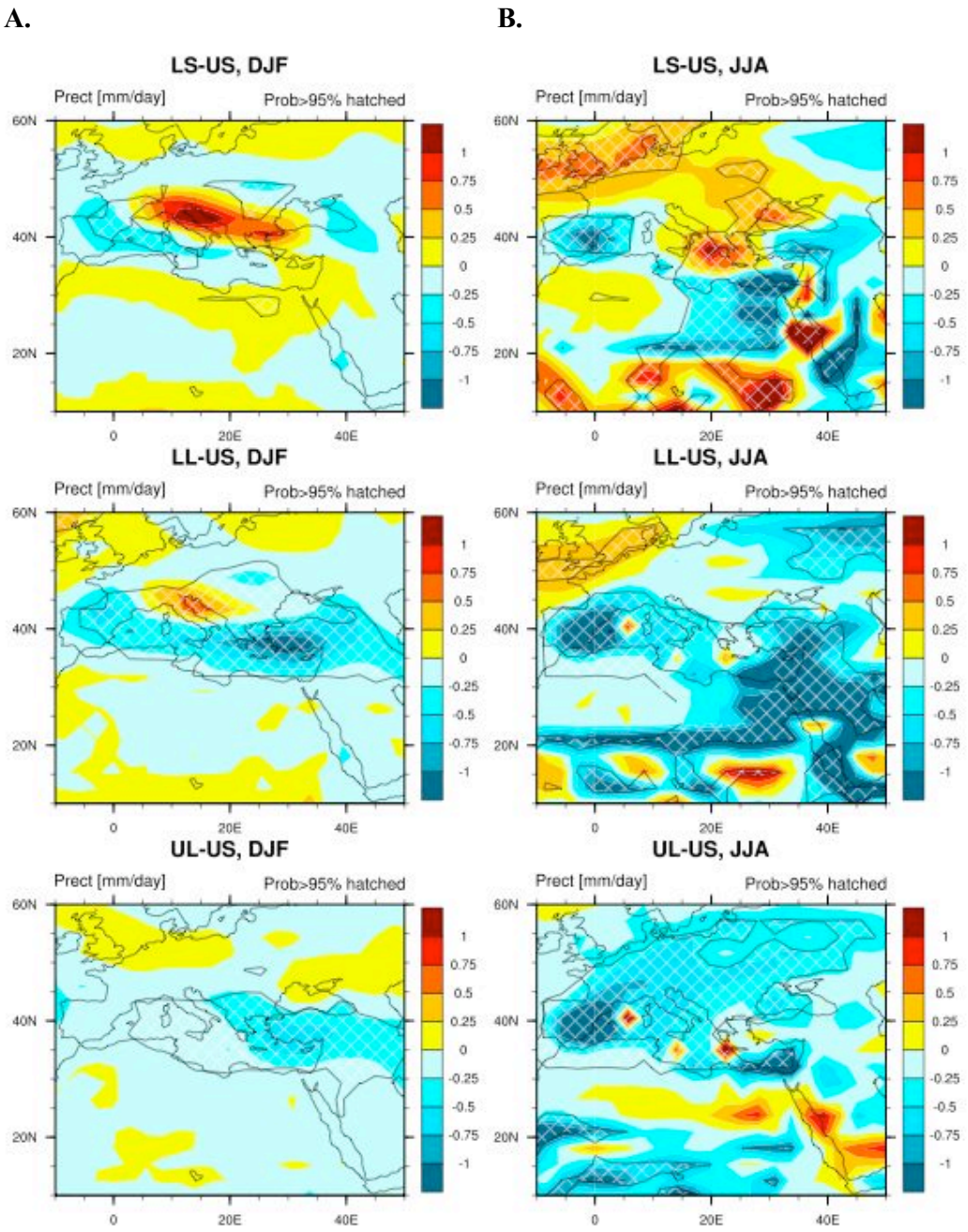


Figure 6. Same as Figure 4 but for total precipitation, including both convective and large scale precipitation output in mm day^{-1} and its zoomed into the Mediterranean region.

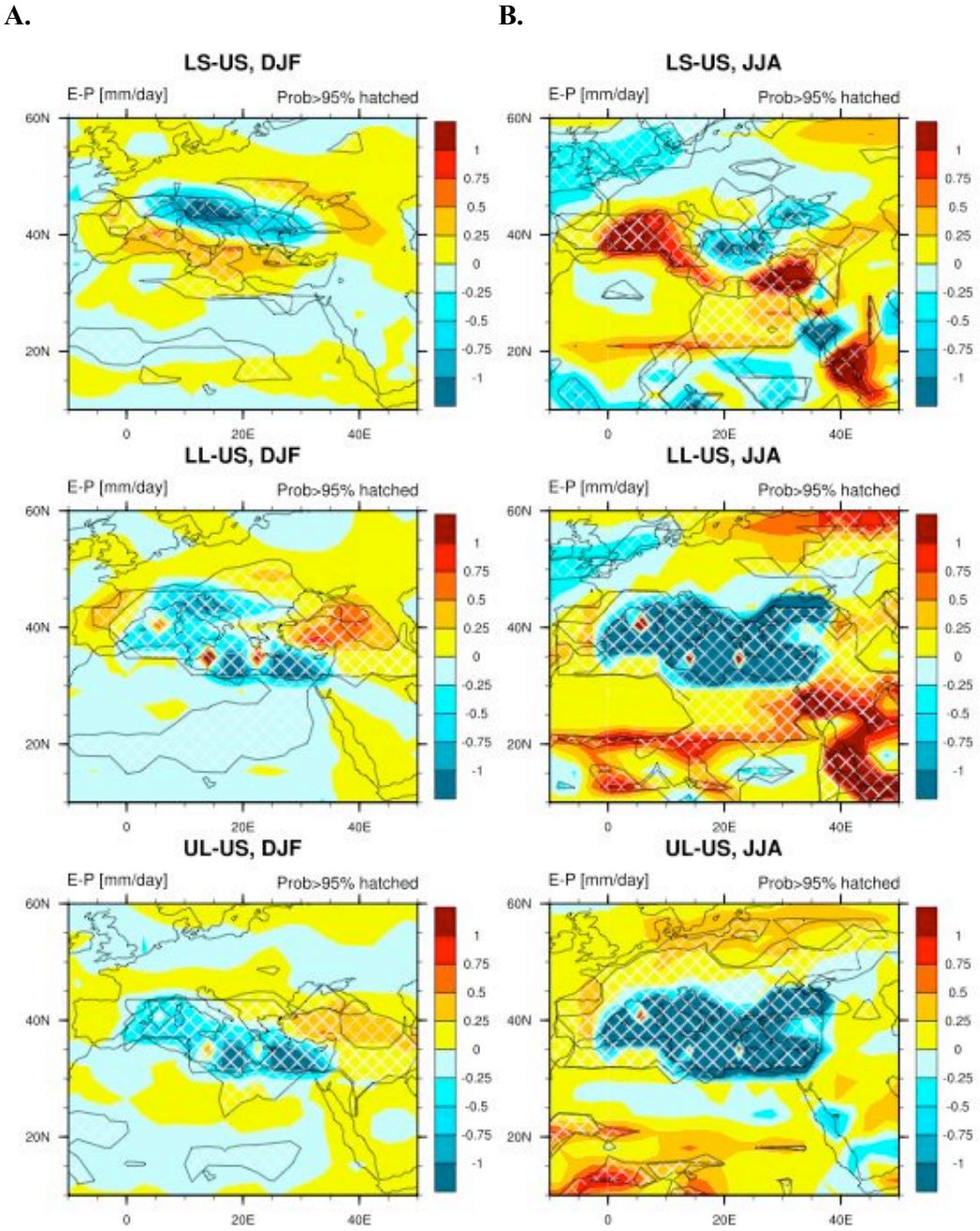


Figure 7. Same as Figure 6 but for the water budget (E-P) in mm day^{-1}.

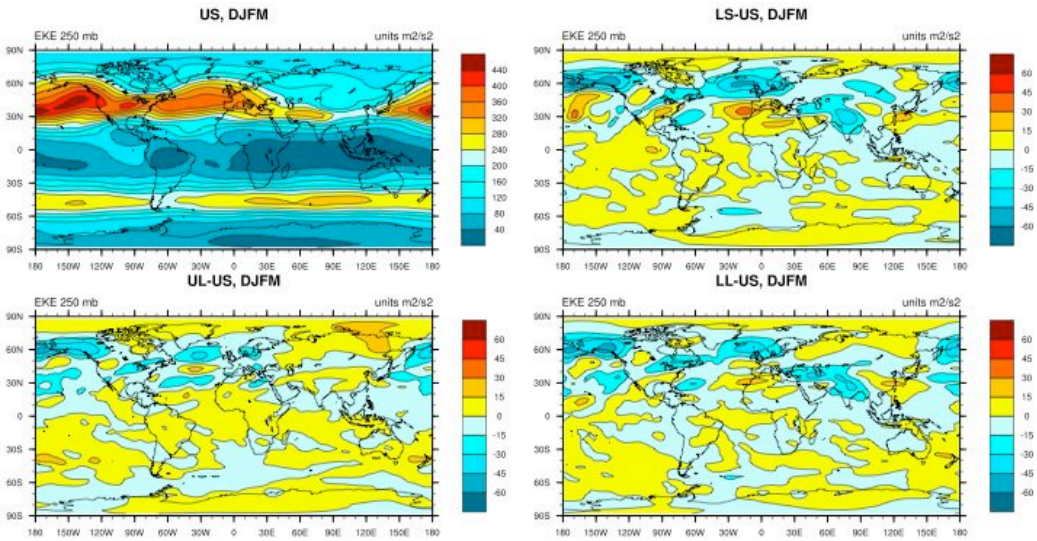


Figure 8. The top left panel shows the climatological Eddy Kinetic Energy in $m^2 s^{-2}$ averaged over the DJFM season. EKE reaches its highest values in the storm track regions. The next three panels show the EKE anomalies averaged over the DJFM season.

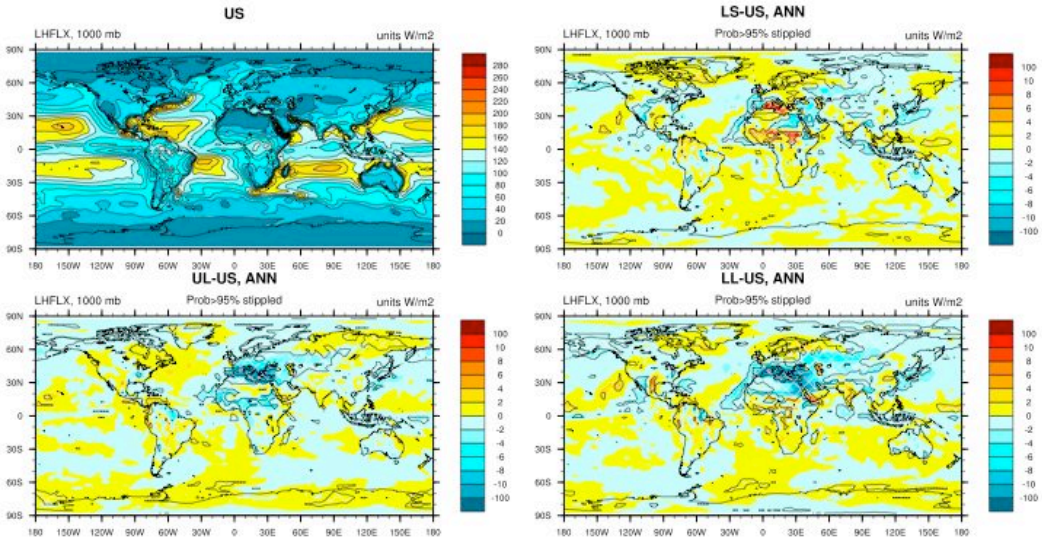


Figure 9. Same as Figure 8 but for the annual mean Latent Heat Flux in W m^{-2}.

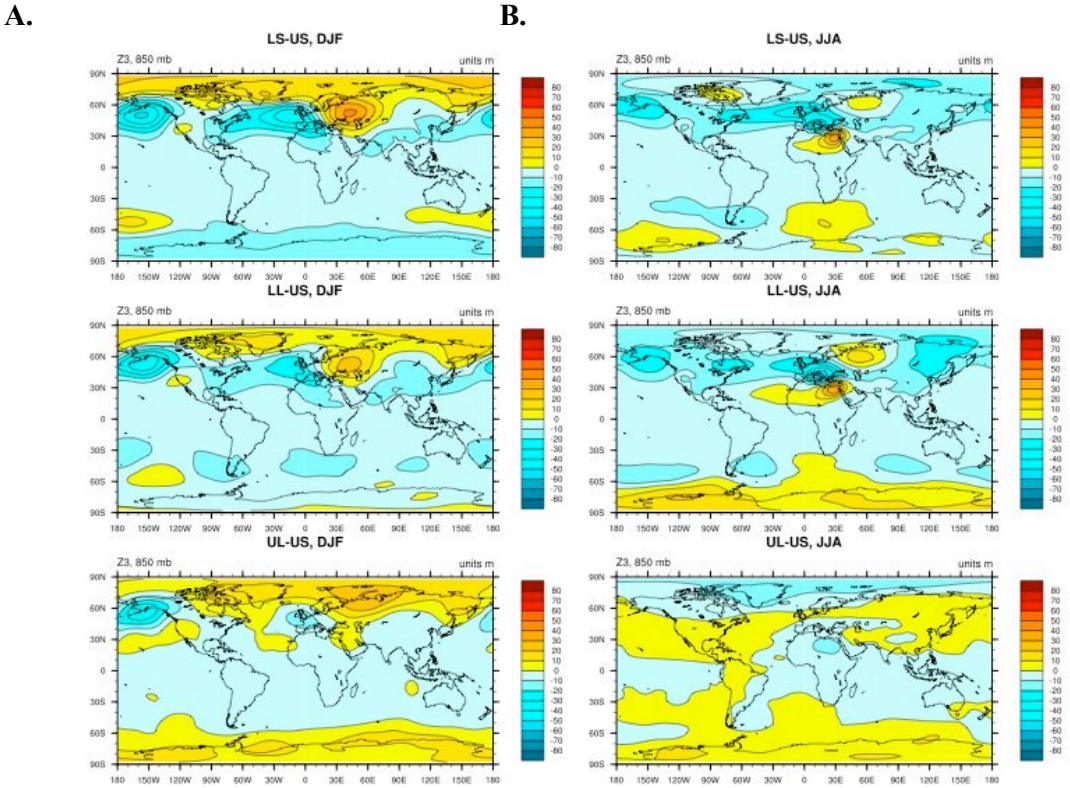


Figure 10. Same as Figure 4 but for geopotential height (Z3) anomalies in meters at 850 hPa.

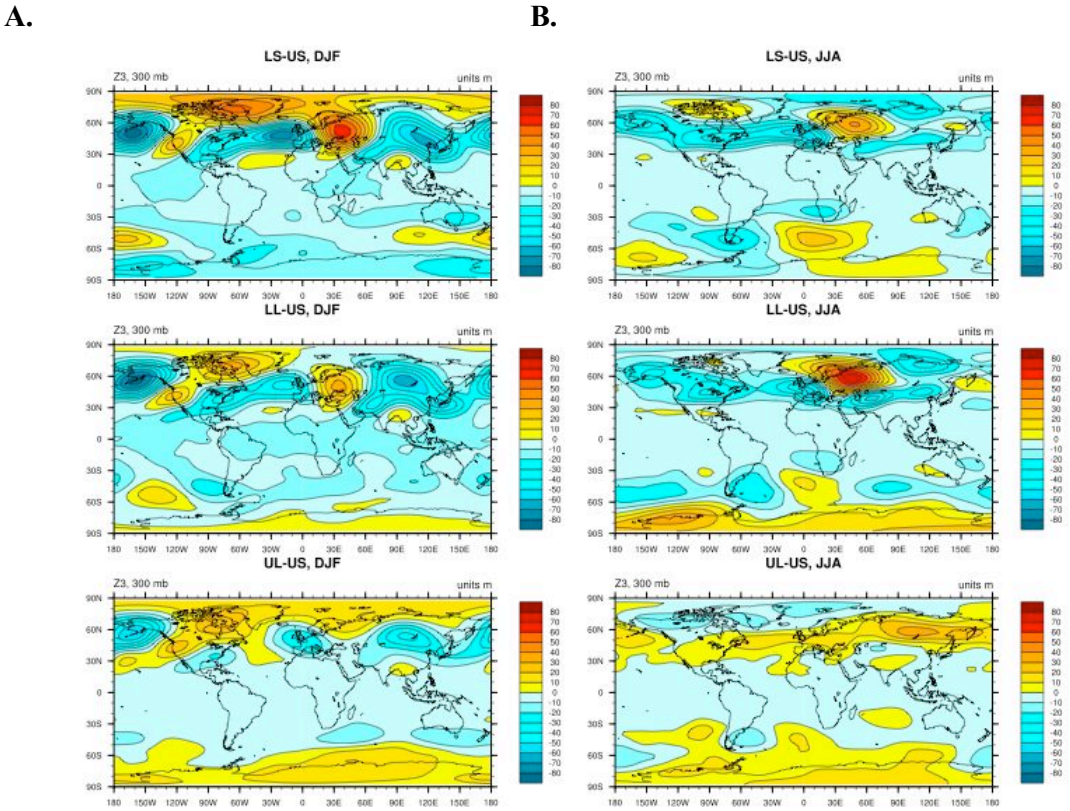


Figure 11. Same as Figure 4 but for geopotential height (Z3) anomalies in meters at 300 hPa.

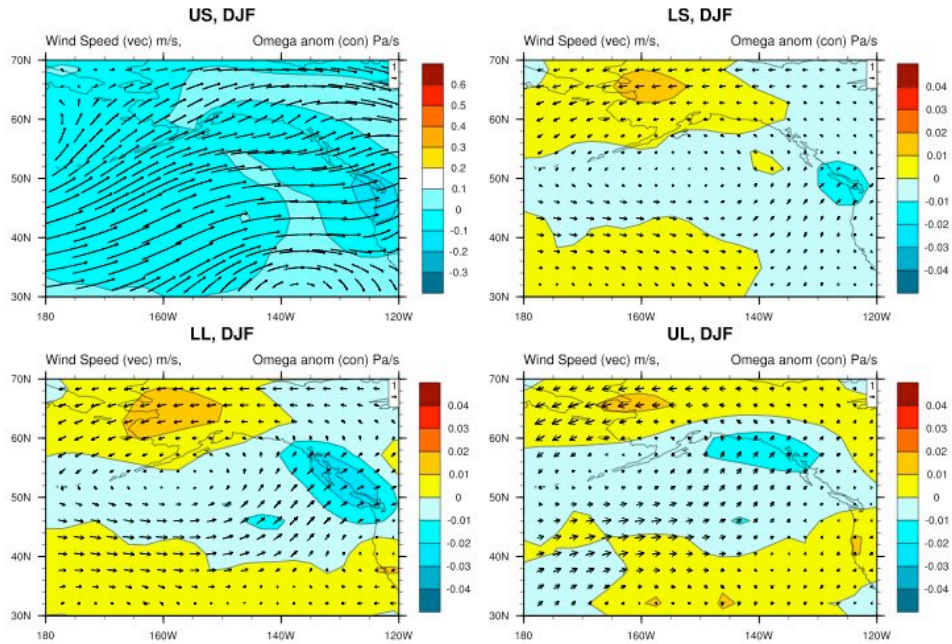


Figure 12. Same as Figure 5 but now centered over the Gulf of Alaska.

A. **B.**

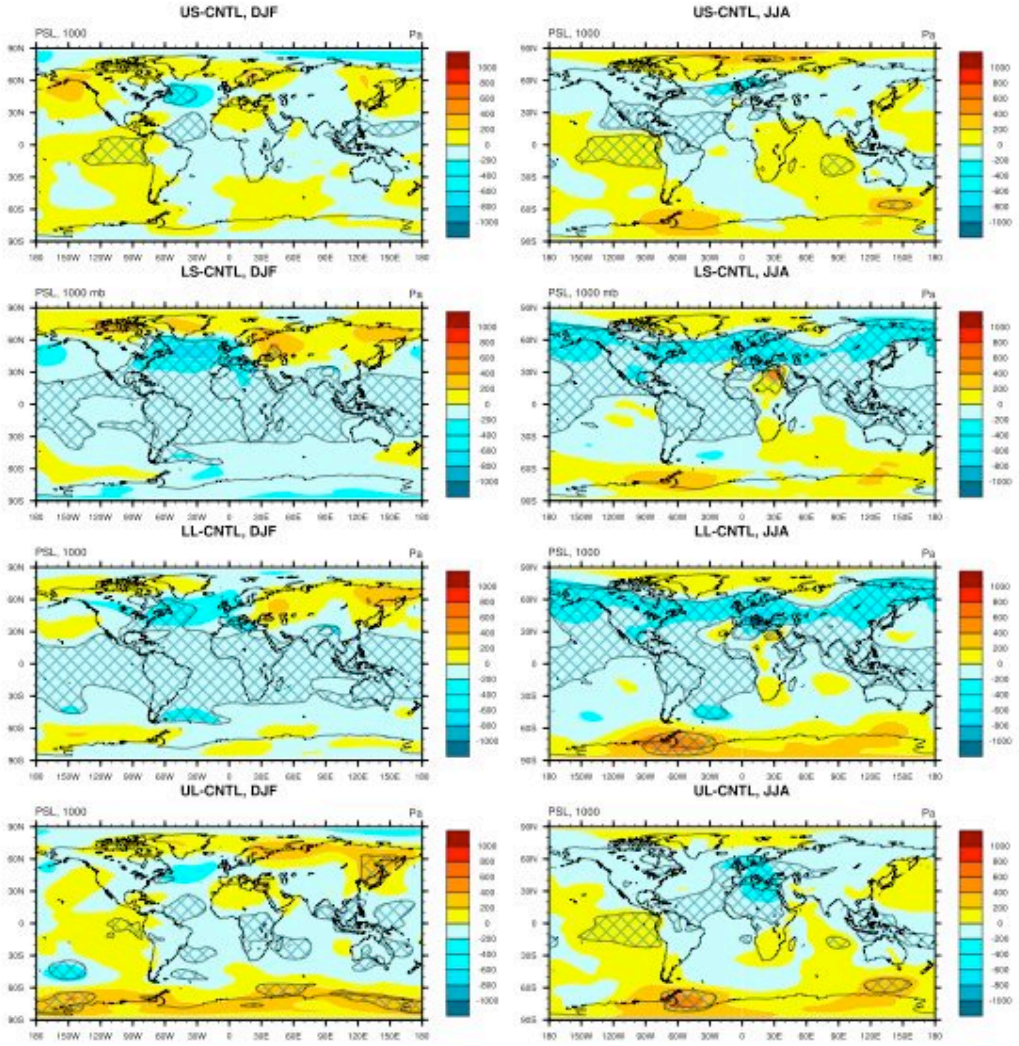


Figure 13. Shows the differences in sea level pressure (PSL) in Pascals between our four simulations and a true CAM3 SOM control run at T42 resolution. Part A shows the DJF differences and Part B shows the JJA differences. Comparing the four middle plots to the top two and bottom two plots illustrate the effect of changing the surface elevation on the atmospheric circulation.

A. **B.**

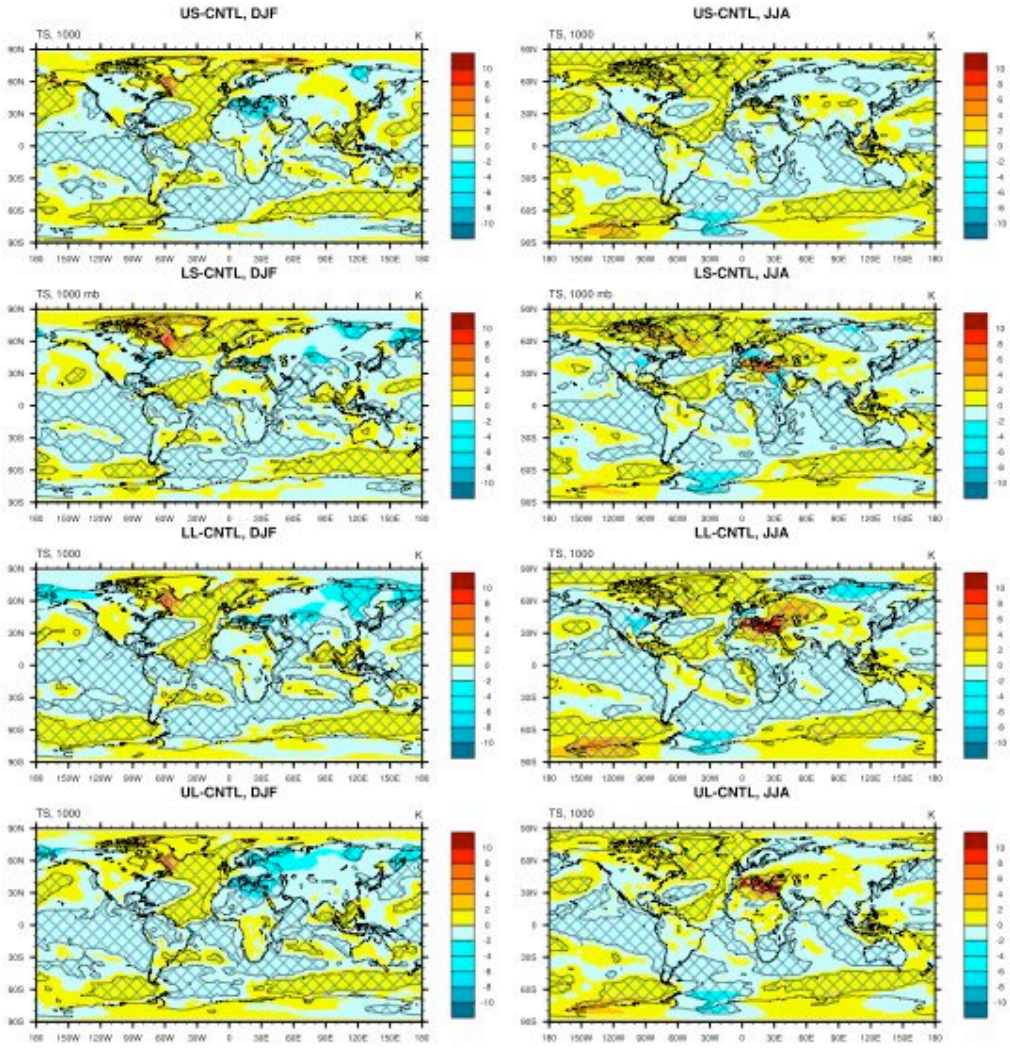


Figure 14. Same as Figure 13 but for surface temperature.

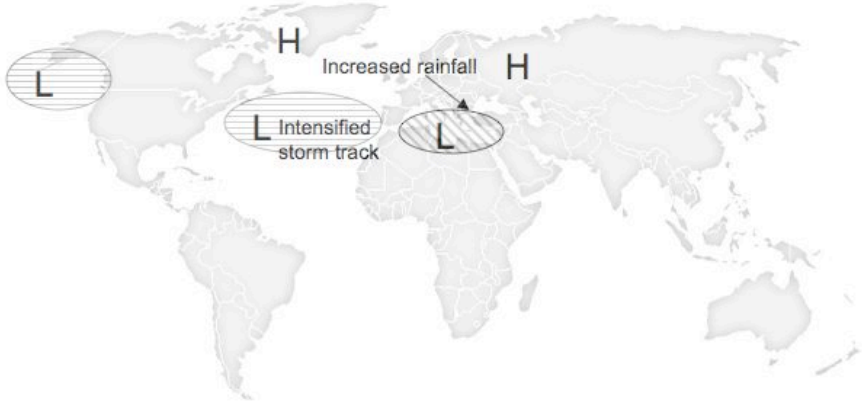


Figure 15. A schematic illustration of the most salient changes that occurred in our simulations of the MSC. The circles with diagonal lines indicate locations where SSTs

warmed. The circles with horizontal lines indicate locations where SSTs cooled. The Hs are regions where the surface pressure rose and the height field was lifted and the Ls indicate regions where the surface pressure fell and the height field was lowered.