

## On the Increased Frequency of Mediterranean Drought

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### ABSTRACT

The land area surrounding the Mediterranean Sea has experienced 10 of the 12 driest winters since 1902 in just the last 20 years. A change in wintertime Mediterranean precipitation toward drier conditions has likely occurred over 1902–2010 whose magnitude cannot be reconciled with internal variability alone. Anthropogenic greenhouse gas and aerosol forcing are key attributable factors for this increased drying, though the external signal explains only half of the drying magnitude. Furthermore, sea surface temperature (SST) forcing during 1902–2010 likely played an important role in the observed Mediterranean drying, and the externally forced drying signal likely also occurs through an SST change signal.

The observed wintertime Mediterranean drying over the last century can be understood in a simple framework of the region's sensitivity to a uniform global ocean warming and to modest changes in the ocean's zonal and meridional SST gradients. Climate models subjected to a uniform  $+0.5^{\circ}\text{C}$  warming of the world oceans induce eastern Mediterranean drying but fail to generate the observed widespread Mediterranean drying pattern. For a  $+0.5^{\circ}\text{C}$  SST warming confined to tropical latitudes only, a dry signal spanning the entire Mediterranean region occurs. The simulated Mediterranean drying intensifies further when the Indian Ocean is warmed  $+0.5^{\circ}\text{C}$  more than the remaining tropical oceans, an enhanced drying signal attributable to a distinctive atmospheric circulation response resembling the positive phase of the North Atlantic Oscillation. The extent to which these mechanisms and the region's overall drying since 1902 reflect similar mechanisms operating in association with external radiative forcing are discussed.

### 1. Introduction

Half of Mediterranean land areas are devoted to agriculture, consuming 60%–80% of water supply (ESPERE 2006). It is in this context that increased drought frequency observed over land areas surrounding the Mediterranean Sea in recent decades (e.g., Mariotti 2010) is of great concern because of the threats posed to the region's food security (Nelson et al. 2009). The region, which accumulates most of its annual precipitation via migratory storms during November–April, is shown in this paper to have experienced 10 of the 12 driest winter seasons since 1902 in just the last 20 years (Fig. 1, top). In concert with this drying trend, which has been a cold season phenomenon to date, precipitation has increased over northern Europe since 1902. The south–north contrast in European precipitation trends over the last

century (Fig. 1, bottom) resembles anomaly patterns occurring on interannual and decadal time scales—often associated with fluctuations of the North Atlantic Oscillation (NAO), the dominant structure of North Atlantic atmospheric variability that is linked to storm track changes over Europe (e.g., Hurrell 1995; Jones et al. 2003; Hurrell and Deser 2009; Mariotti and Dell'Aquila 2011). Also, such a south–north contrast in precipitation trends describes the pattern of climate change in projections toward the end of the twenty-first century in models forced by scenarios of increasing anthropogenic greenhouse gases conducted as part of the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment (Ulbrich et al. 2006; Solomon et al. 2007; Mariotti et al. 2008). By the latter decades of the twenty-first century, the Mediterranean region is projected to experience the greatest drying (as a percentage of climatological rainfall) among 26 regions across the globe (Giorgi 2006).

The latter acuteness in IPCC model projections of anthropogenic impacts has earned the Mediterranean region the title of major “climate change hot spot”

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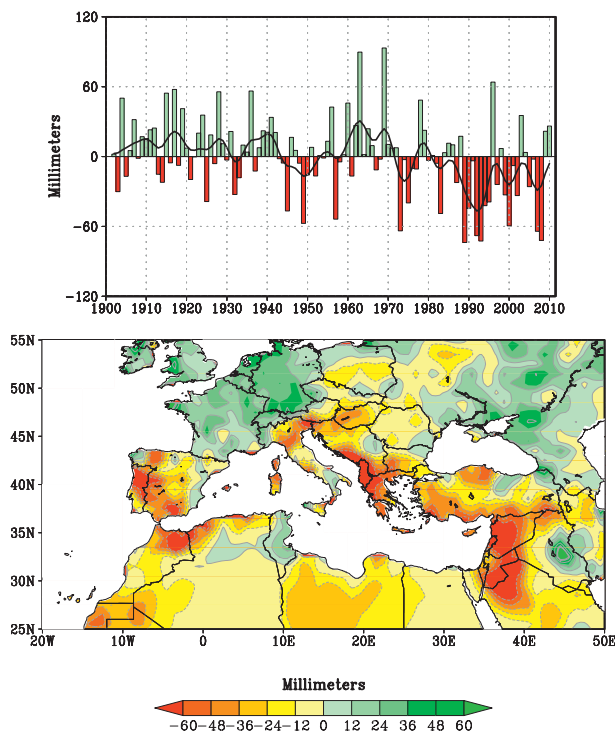


FIG. 1. (top) Observed time series of Mediterranean (30°–45°N; 10°W–40°E) cold season (November–April) precipitation for the period 1902–2010 and (bottom) the observed change in cold season precipitation for the period 1971–2010 minus 1902–70. Anomalies (mm) are relative to the 1902–2010 period. Solid curve is the smoothed precipitation time series using a nine-point Gaussian filter. Data are from the GPCC.

(Giorgi 2006). It is believed that changes in zonal mean atmospheric circulation would become an effective mechanism for inducing a drying of lower midlatitudes, including the Mediterranean region, in response to increasing greenhouse gas (GHG) concentrations (e.g., Seager et al. 2007, 2010). Such drying would occur mainly through a poleward shift of subtropical dry zones due to a poleward expansion of the Hadley cell (e.g., Lu et al. 2007) and a poleward shift of the storm tracks (e.g., Yin 2005; Previdi and Liepert 2007).

This paper explores possible causes for increased Mediterranean drought observed during recent decades, the results of which contribute to the current understanding of the physical processes exposing the Mediterranean region as one of heightened sensitivity to anthropogenic greenhouse gas forcing. To understand the mechanisms associated with recent observed drying, we focus on the responsiveness of the region's wintertime precipitation to patterns of sea surface temperature (SST) forcing that describe principal features of the ocean's change during the last century. Previous studies have shown that global SSTs exert an important

controlling effect on North Atlantic/European climate via a sensitivity of the wintertime NAO (e.g., Mehta et al. 2000; Robertson et al. 2000; Hoerling et al. 2001; Sutton et al. 2000); in particular, a trend toward a positive phase of the NAO during 1950–99 was found to be consistent with an atmospheric response pattern forced by Indian Ocean warming (Bader and Latif 2003; Hurrell et al. 2004; Hoerling et al. 2004; Li et al. 2006). Given the strong external radiative contribution to Indian Ocean warming in recent decades (e.g., Knutson et al. 2006), these studies have advanced a theory that anthropogenic forcing may have already contributed to recent North Atlantic and Mediterranean climate trends. Such a notion appears to be consistent with the subsequent analyses of Mariotti et al. (2008) who document a decline in Mediterranean annual precipitation during the twentieth century in the Coupled Model Intercomparison Project phase 3 (CMIP3) simulations (at the rate of  $-0.007 \text{ mm day}^{-1} \text{ decade}^{-1}$ ), followed by more rapid drying after 2020.

However, there are certain facts that a hypothesis of climate change forcing of Mediterranean wintertime drying has failed to explain. Foremost, whereas most coupled models subjected to twentieth-century time-varying external radiative forcing yield Indian Ocean warming, there is not a consistent North Atlantic atmospheric circulation response resembling the positive NAO index in those models (e.g., Gillett 2003; Osborn 2011). Here we will demonstrate that it is not the Indian Ocean warming per se to which the NAO is primarily sensitive, but the difference between warming rates of the Indian Ocean and the remainder of the tropical oceans. Furthermore, updated observational time series of NAO indices reveal that a trend toward the positive polarity of the NAO, which was particularly strong when diagnosed for the 1950–99 period as done in most prior studies, is considerably weaker when using additional data through 2010. The question has been raised, therefore, whether the observed variations in the winter NAO index can be explained as internally generated atmospheric variability alone (Wunsch 1999; Osborn 2011). Indeed, analysis presented herein suggests only a small, statistically insignificant, upward trend of the NAO during 1902–2010, which fails to explain the considerable wintertime drying of the Mediterranean during this period.

It is also not known whether, or to what extent, the zonal mean circulation has changed since the beginning of the twentieth century, in particular if the relationship between tropical and midlatitude zones has changed through which increased Mediterranean drought might emerge. Davis and Rosenlof (2012) illustrate the sensitivity of recent trends in the width of the tropical belt to the use of various subjective definitions such as the tropopause height or a threshold value of outgoing

longwave radiation. For one particular indicator of the width of the tropical belt, namely, the Hadley cell extent, they find a statistically significant poleward expansion in three of four reanalysis products for the period 1979–2009. Other indicators of the width of the tropics, including ones based on observations, do not show significant change. For instance, there has been no detected change in the zonal averages of land precipitation during 1925–99 within the extratropical band 30°–50°N (Zhang et al. 2007), though this may result from the cancellation of strong regional signals of precipitation change that have opposite signs.

In the current study, we employ global methods involving climate models subjected to external radiative and surface boundary forcings so as to understand drought spanning the region that includes all Mediterranean land areas. Section 2 describes the observational datasets and model simulations, many of which have been used in prior climate studies. New suites of atmospheric model simulations are also conducted that use idealized patterns of SST anomalies. These, also described in section 2, facilitate reconciliation between observed increases in drought frequency, the sensitivity to external radiative forcing, and the role of oceanic conditions and atmospheric circulation patterns.

Results are given in section 3, where we present evidence that a change in the region's climate has been detected and that it is unlikely that the observed November–April Mediterranean drying since 1902 occurred due to internal variability alone. Diagnosis of the CMIP3 coupled models reveals that this detected change toward drier conditions is attributable, in part, to the Mediterranean region's sensitivity to time-evolving external radiative forcing. The amplitude of the externally forced, area-averaged drying signal is roughly one-half the magnitude of the observed drying during 1902–2010, indicating that other processes likely also contributed. Section 3 presents analyses of atmospheric GCM simulations forced with observed SST and sea ice variability since 1902 but using fixed external radiative forcing [so-called Atmospheric Model Intercomparison Project (AMIP) simulations]. These reveal that SST forcing was likely an important factor in the observed drying and, furthermore, that the externally forced drying signal likely also occurs through an impact of an SST change. Section 4 explores the sensitivity of Mediterranean climate to idealized patterns of SST forcing that depict gross global features of the CMIP3 and observed patterns of SST change since 1902. It is found that a globally uniform +0.5°C warming of the world oceans produces drying over the eastern Mediterranean but fails to produce widespread drying over all Mediterranean land areas. A

larger amplitude and more expansive drying that includes all Mediterranean land areas emerges when tropical SST warming is greater than extratropical SST warming (a meridional gradient). The magnitude of drying increases yet further in the presence of an oceanic condition in which the tropical Indian Ocean SSTs are warmed more than the rest of the tropical oceans (a zonal gradient). These rainfall sensitivities are explained in the context of atmospheric circulation responses to the SST-forcing scenarios. Summary comments appear in section 5, and we offer a holistic view of physical mechanisms contributing to increased wintertime Mediterranean drought during recent decades including processes by which the Mediterranean region likely responds to anthropogenic greenhouse forcing.

## 2. Data and methods

### *a. Observational data and analysis period*

Global monthly SST data is based on the 5°-gridded NOAA product (Smith and Reynolds 2005). We also analyze monthly 500-hPa geopotential heights for the period 1902–2008 based on the new twentieth-century reanalysis data (Compo et al. 2011).

Several different gridded analyses of monthly precipitation are used. The Global Precipitation Climatology Centre (GPCC) analysis (Rudolf and Schneider 2005) is gridded at 1° resolution covering the period 1901–2010 and was used to construct Fig. 1. The University of Delaware data (UDEL) (Legates and Willmott 1990) is gridded at 0.5° resolution and covers 1900–2008. A third dataset is the Hadley Center's Climate Research Unit (CRU) analysis (HadCRU) (Hulme et al. 1998) also gridded at 0.5° resolution covering 1901–2006. And finally, the Global Historical Climate Network (GHCN) (Vose et al. 1992) data gridded at 5° resolution covering the period 1880–2010.

Figure 2 (top) compares Mediterranean area-averaged precipitation time series for the above four analyses. They broadly agree in portraying a dry period after 1970, which contrasts with mostly above normal precipitation during prior decades. The analyses exhibit strongest agreement during 1930–70, and the larger spread among analyses during both the earlier and later decades is likely related to sparser data coverage (not shown). The calculated change in area-averaged Mediterranean precipitation (1971–present versus 1902–70) as a percentage change of the November to April climatology is –6.8%, –10.1%, –5.1%, and –5.1% for the GPCC, UDEL, CRU, and National Climatic Data Center (NCDC) analyses, respectively. The four dataset averaged decline is –6.8%.

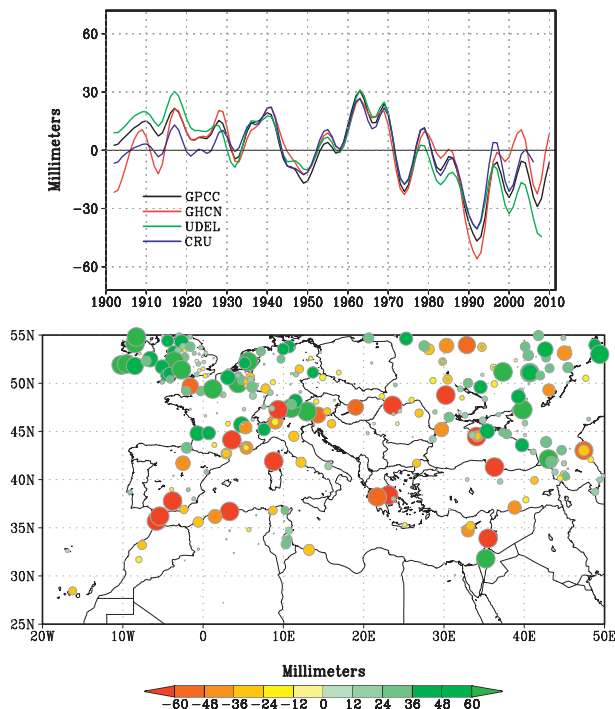


FIG. 2. (top) Time series of Mediterranean ( $30^{\circ}$ – $45^{\circ}$ N,  $10^{\circ}$ W– $40^{\circ}$ E) cold season (November–April) precipitation for the period 1902–2010 for four observed datasets: GPCC, GHCN, UDEL, and CRU. Anomalies (mm) are referenced to the 1902–2010 period and are smoothed with a nine-point Gaussian filter. (bottom) A station-based analysis of the change in cold season precipitation for the period 1971–2010 minus 1902–70 for stations are extracted from the GHCN (monthly) and GHCN-d (daily) datasets.

The bottom panel of Fig. 2 illustrates the density of reporting stations that have a minimum of 80% available observations during the 109-yr period. There are considerable monitoring gaps (e.g., the northwest Iberian Peninsula, southeast Europe, and western Turkey) indicating that caution should be exercised when assessing subregional structures of rainfall differences in the gridded analyses. Nonetheless, the larger-scale regional pattern of precipitation differences (1971–2010 versus 1902–70) at the GHCN stations supports the realism of patterns constructed using the gridded analyses (cf. Fig. 1).

### b. Climate model simulations with realistic forcing

Two configurations of climate model simulations are used to assess factors that may have influenced Mediterranean precipitation variability since 1902: atmospheric general circulation models (AMIP) and coupled ocean–atmosphere general circulation models (CMIP3). For the former, a total of three different models were available, each subjected to specified monthly varying

observed global SSTs but using climatological values for the atmospheric composition.<sup>1</sup> The total ensemble size of AMIP runs is 40 members. For CMIP3 simulations, a total of 22 different models were utilized, each subjected to specified monthly variations in greenhouse gases, aerosols, and half also including time-varying solar irradiance and the radiative effects of volcanic activity for 1880–1999. The IPCC Special Report on Emissions Scenarios (SRES) A1B (Alley 2007) is used thereafter. Our analysis uses single runs from each of the modeling centers yielding a total ensemble of 22 members, with the model data accessed from the Program for Climate Model Diagnosis and Intercomparison (PCMDI) archive as part of the Coupled Model Intercomparison Project (CMIP3) (Meehl et al. 2007). The externally forced (greenhouse gas, aerosol, solar, and volcanic) signal in precipitation is estimated by averaging the multimodel CMIP3 ensemble members, whereas the SST-forced signal is estimated by averaging the multimodel AMIP ensemble members.

Unforced control integrations of CMIP3 coupled models are also diagnosed. Most modeling centers generated roughly 300-yr simulations using climatological well-mixed greenhouse gases associated with pre-1900 conditions. We calculate the statistics of 109-yr precipitation change (equivalent in duration to the span of 1902–2010) using 19 of the models having long integrations to estimate the influence of internal coupled ocean–atmosphere noise.

### c. Climate model simulations with idealized SST forcing

An additional suite of atmospheric climate model simulations is conducted that specify SST anomalies representing various idealizations of the observed and CMIP3 simulated SST changes since 1902. The change in SSTs (1971–2010 minus 1902–70) is shown in Fig. 3 based on observations and the CMIP3 ensemble average. The outstanding feature of the SST difference is an overall warming of the oceans with a generally greater tropical SST warming than that occurring over the NH extratropics. Though of weaker magnitude, there is also an enhanced zonal gradient of tropical SST warming between the Indian and Pacific Ocean. In general, the observed SST differences have stronger meridional

<sup>1</sup> The three atmospheric models used are the NCAR Community Climate Model (CCM3) (16 members, Kiehl et al. 1998), the NASA Seasonal-to-Interannual Prediction Project (NSIPP) model (14 members, Schubert et al. 2004), and the Geophysical Fluid Dynamics Laboratory Atmospheric Model version 2.1 (AM2.1) (10 members, Delworth et al. 2006).



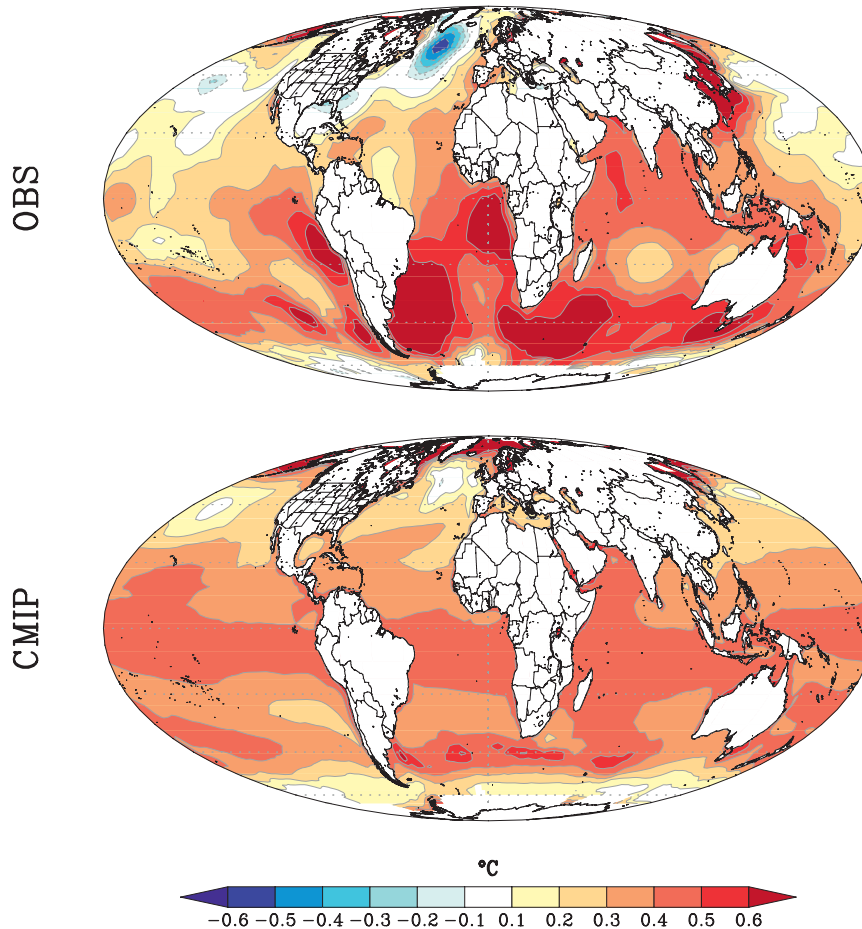


FIG. 3. Cold season (November–April) sea surface temperature departures ( $^{\circ}\text{C}$ ) for the period 1971–2010 minus 1902–70: (top) observed analysis from the NOAA merged gridded dataset and (bottom) multimodel CMIP3 ensemble sea surface temperatures based on 22 IPCC models.

contrast between the tropics and NH extratropics and also a stronger zonal contrast between the Indian Ocean and the tropical Pacific Ocean. This comparison motivates a set of additional AGCM experiments using idealized SSTs forcings including (i) uniform  $+0.5^{\circ}\text{C}$  warming of global SSTs, (ii) uniform  $+0.5^{\circ}\text{C}$  warming of tropical ( $30^{\circ}\text{N}$ – $30^{\circ}\text{S}$ ) SSTs only, thereby introducing a simple meridional gradient, and (iii) same as for (ii) but an additional  $0.5^{\circ}\text{C}$  warming of the tropical Indian Ocean thereby introducing a simple zonal gradient. We employ two models (GFDL and NASA) of the three atmospheric models used in AMIP mode and generate respectively a 30- and 50-member ensemble subjected to the three SST forcings. The atmospheric sensitivity to the specified SST changes in these runs is determined by comparing with 50-yr control integrations of each model that used a 1971–2000 climatologically averaged, seasonally varying SST.

### 3. Mediterranean precipitation variability during 1902–2010

#### a. Detection of change

The time series of area-averaged land precipitation ( $30^{\circ}$ – $45^{\circ}\text{N}$ ,  $10^{\circ}\text{W}$ – $40^{\circ}\text{E}$ ) shows a different character for precipitation variability after 1970 when drought was more prevalent relative to prior decades when precipitation was consistently more abundant (Fig. 1, top). The visual impression of two different precipitation regimes during this instrumental record motivates a comparison of the spatial pattern of precipitation differences between the two epochs (1971–2010 minus 1902–70). The results reveal widespread drying across North Africa and southern Europe extending from the Atlantic coast to the Middle East (Fig. 1, bottom).

A true change in the region's observed precipitation has *likely* occurred in a statistical sense during the last

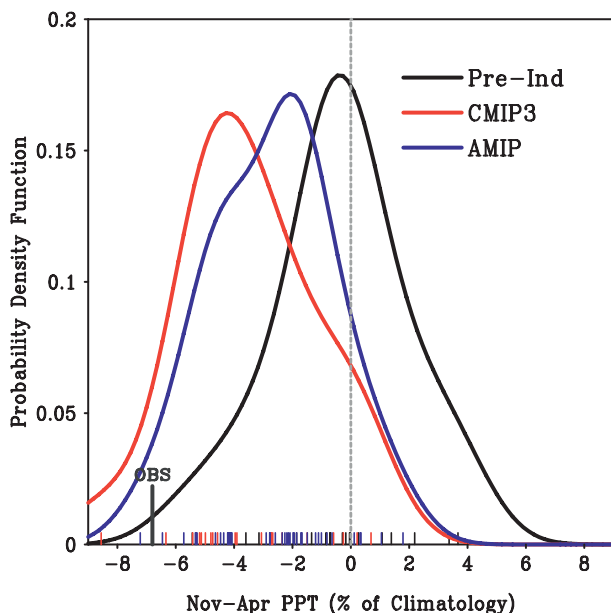


FIG. 4. Probability distribution functions for the 1971–2010 minus 1902–70 period anomalies of cold season precipitation over the Mediterranean region as a percentage of climatology. The red curve is based on the analysis of 22 IPCC models, while the blue curve is based on 40 AMIP simulations of the twentieth century. The solid black curve is drawn from the analysis of 109-yr-long samples of unforced integrations from 19 IPCC simulations using preindustrial concentrations of greenhouse gases. The observed precipitation index decline, based on the average of four analyses ( $-6.8\%$ ), is shown by the gray tick mark to the left; colored tick marks indicate individual ensemble members.

century, as revealed by comparing the observed trend with those calculated from simulations of preindustrial unforced coupled models. We analyze continuous 109-yr integrations for each of the 19 CMIP3 models and calculate epochal differences of Mediterranean winter season precipitation in a manner analogous to what was done with the observed record. The results, plotted as a probability distribution function (PDF) in Fig. 4 (black curve), indicate that no single sample of 109-yr model time series yields a drying as strong as that observed during the twentieth century. We estimate a less than 1% probability that Mediterranean drying as strong as observed since 1902 could occur in unforced coupled models, based on analysis of the area under the curve of the non-parametric fit to the model results. This is based on using a  $-6.8\%$  decline, the average of the four analyses. We conclude it is *unlikely* that the observed increase in drought frequency over the Mediterranean region can be attributed to internal variability alone. Our confidence language reflects the uncertainty in knowing the amplitude of the true precipitation decline and the presence of model biases in estimating the internal variability of the observed system.

### b. Attribution of change

We find that the recent increase in Mediterranean drought is consistent with the region's sensitivity to anthropogenic greenhouse gas and aerosol forcing, though the amplitude of the externally forced drying signal is only one-half the amplitude of observed drying. Figure 5 shows the time series of Mediterranean cold season precipitation derived from the average of the 22 externally forced CMIP3 models (top, left), together with the spatial map of the ensemble averaged twentieth-century precipitation change (bottom left). The temporally evolving signal describes increased dryness in recent decades, consistent with Mariotti et al. (2008). The precipitation decline is not gradual; the time series is noteworthy for a sudden shift to a sustained dry signal after the 1970s, a temporal behavior analogous to the observed time series (cf. Fig. 1). The spatial pattern of precipitation change in the CMIP3 ensemble consisting of a meridional dipole of North African/southern European decline and northern European increase is also similar to observations.

We find a substantial increase in the probability of a Mediterranean drying trend in the externally forced simulations relative to their likelihood of occurrence in preindustrial unforced counterparts. A second PDF plotted in Fig. 4 (red curve) shows the Mediterranean area-averaged precipitation change (1971–2010 minus 1902–70) simulated in each of the 22 CMIP3 models and illustrates a reproducible drying; virtually all of the model simulations yield a reduction in precipitation during the latter decades of the twentieth century, although only one CMIP member yields a drying greater than observed. The mean values of the two PDFs are found to be statistically different at a 99.9% confidence level. When comparing the areas under the curves, we estimate a 10-fold increase in the relative probability of drying occurrences in the CMIP3 twentieth-century simulations versus unforced simulations that is as large as the observed drying. The detected change in Mediterranean wintertime precipitation is thus attributable, in part, to the time varying external radiative forcing.

Anthropogenic greenhouse gases affect Mediterranean precipitation principally via a sensitivity to the contemporaneous response of the oceans to external radiative forcing, rather than by an influence of local radiative forcing itself. This assessment is drawn from diagnosis of the precipitation time series derived from the AMIP simulations subjected only to the observed SST and sea ice variability (Fig. 5, top right) and also the spatial map of the AMIP ensemble-averaged twentieth-century precipitation change (Fig. 5, bottom right). Further supporting evidence is given in section 4 based on additional simulations using idealized features of the CMIP3 and observed SST change patterns. The ocean-forced signal

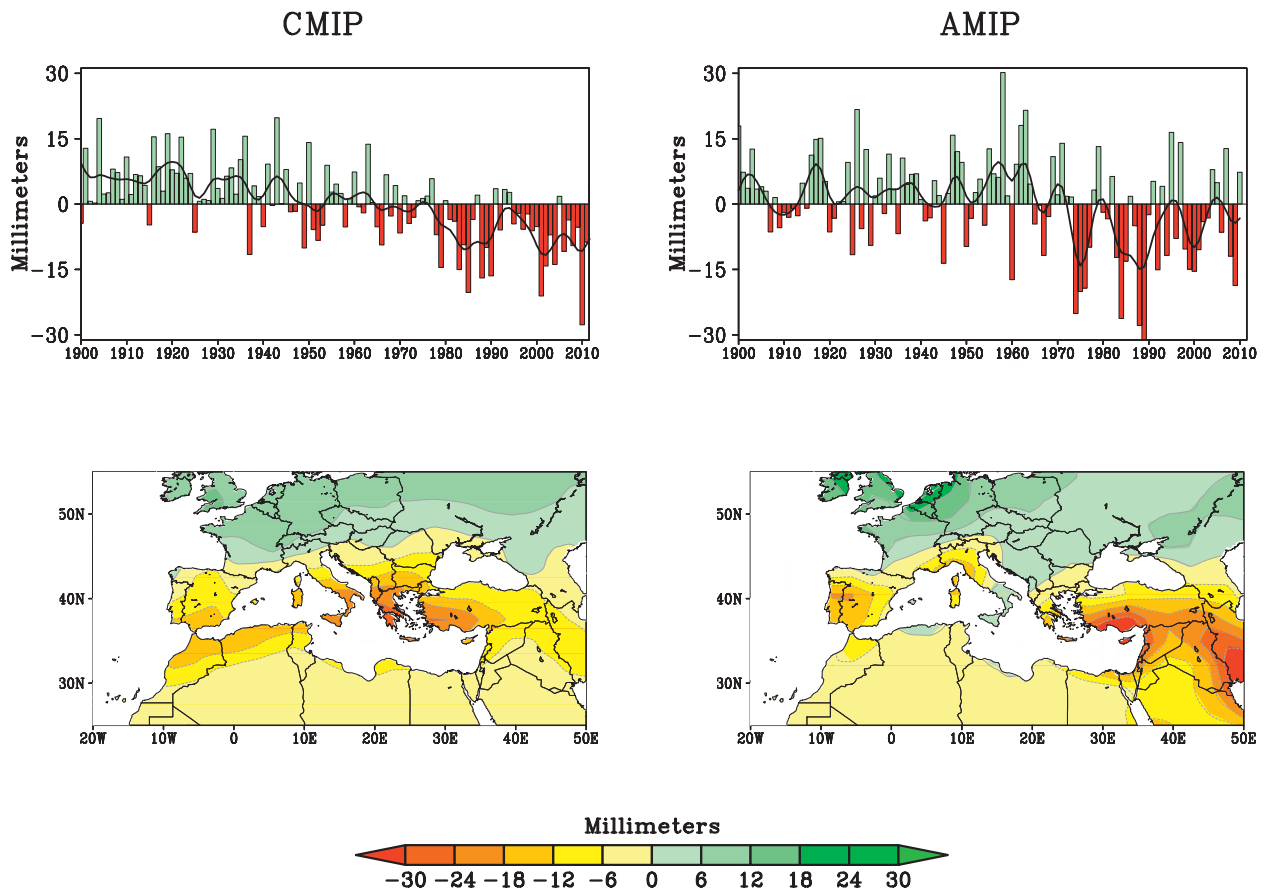


FIG. 5. Simulated change in cold season precipitation (mm) over the Mediterranean region as in Fig. 1 based on the ensemble average (left) of 22 IPCC models and (right) of 40 AMIP twentieth-century runs. The difference plots in the lower panels are for the period 1971–2010 minus 1902–70.

in the AMIP experiments is one of dryness during much of the post-1970 period, analogous to the period of dryness simulated in CMIP3 in response to anthropogenic greenhouse gas and aerosol forcing and also analogous to the observed post-1970 increase in drought frequency. The spatial pattern of the ocean signal describes a dipole pattern of Mediterranean dry and north European wet conditions, largely reproducing the pattern of the anthropogenic greenhouse gas and aerosol signal and highly congruent with the observed change pattern (cf. Fig. 1).

The observed Mediterranean drying cannot be readily attributed to an observed change in behavior of the North Atlantic Oscillation. First, a NAO index time series for 1902–2010 fails to exhibit trendlike behavior analogous to that seen for Mediterranean rainfall and, instead, exhibits only a small, and statistically insignificant, upward trend during the 109-yr period (Fig. 6, top). Second, the NAO-related signal of Mediterranean precipitation change post-1970, as estimated from a linear regression analysis, while showing drying over most of the region, has much weaker amplitudes than

observed. Thus, it fails to explain the severity of overall drying and also does not account for several regional trend features including strong drying over the eastern Mediterranean and a coincident increase in central and northern European precipitation. We note furthermore that the NAO behavior post-1970 is not materially different than that before 1970 and certainly has not changed in the dramatic manner characterizing the temporal evolution of observed Mediterranean rainfall.

It is interesting to note that our initial selection of these two epochs before and after 1970 was based purely on inspection of the observed time series of Mediterranean precipitation without any consideration of possible changes in forcings. The CMIP3 and AMIP simulations substantiate, however, that a temporal change in forcing has indeed occurred so as to distinguish pre- and post-1970, and that the key forcing involves the oceans.

A third PDF, shown in Fig. 4, illustrates the statistics of precipitation trends occurring in each of the AMIP simulations (blue curve). We again find that the mean value of the PDF associated with observed ocean forcing during

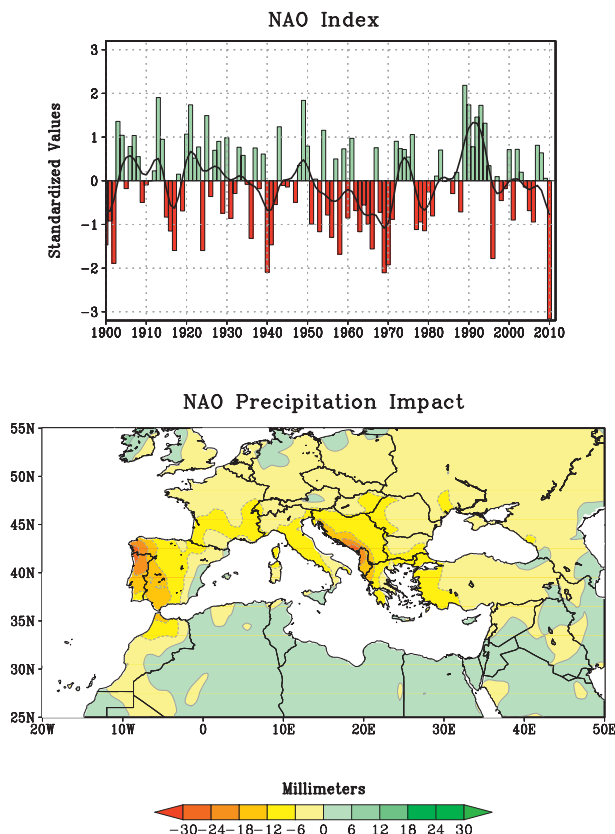


FIG. 6. Time series of an index of the observed NAO during (top) the cold season (November–April) for the period 1902–2010 and (bottom) the change in cold season precipitation for the period 1971–2010 minus 1902–70 that is linearly related to the change in the NAO index. The NAO index is the principal component time series of the leading empirical orthogonal function of monthly 500-hPa heights during November–April over the region 20°–90°N, 120°W–60°E. Data is the twentieth-century reanalysis, and solid curve is the smoothed NAO time series using a nine-point Gaussian filter. The precipitation impact (mm) of the NAO change is derived by regressing November–April precipitation onto the NAO index, and then scaling by the mean difference of the NAO index between the two periods (+0.2 standardized units).

1902–2010 is statistically different from that occurring in the preindustrial unforced coupled models at 99% confidence. However, the CMIP3 and AMIP PDFs are found not to be different at 90% confidence. We note also that, as with the CMIP results, the SST-forced drying signal is weaker than the observed drying, and only one of the AMIP members yields drying greater than was observed.

#### 4. Mediterranean climate sensitivity to idealized SST forcing

##### a. Precipitation sensitivity

The results of section 3 indicate that SST forcing has likely been an important contributor to recent increases

in Mediterranean drought. Furthermore, given the similarity between the CMIP3 and AMIP low frequency Mediterranean precipitation time series over the last century, the physics of the externally forced drying signal primarily involves forcing from the oceans.

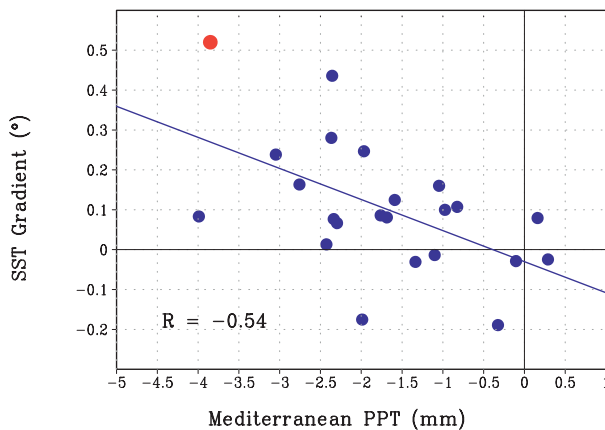
We thus assess the sensitivity of Mediterranean wintertime precipitation to SST scenarios that reflect three gross elements of the observed global SST change during the last century (see Fig. 3): an overall global ocean warming, a greater warming of tropical oceans relative to the Northern Hemisphere (NH) extratropical oceans, and a greater warming of the tropical Indian Ocean relative to the tropical Pacific Ocean. We use a +0.5°C SST perturbation to provide a reasonable rendering of SST changes observed during the last century, although the +0.5°C enhanced warming of the Indian Ocean relative to the tropical Pacific is a factor of 4 greater than the observed gradient change since 1902, which should be kept in mind when using these additional model results to understand quantitative aspects of the AMIP and CMIP simulations. We have conducted experiments with two different atmospheric models (see section 2), the results of which are found to be highly consistent with each other, and below we show only the results from the GFDL model.

Before discussing the results of the forced AGCM experiments, it is useful to first examine the statistical relationship between these various attributes of SST change and Mediterranean precipitation occurring in the separate models of the twentieth-century externally forced CMIP3 ensemble. We had previously shown (Fig. 3) that, whereas most oceans experience a warming, neither the observed SST change nor the SST response to anthropogenic greenhouse gas forcing are spatially uniform patterns. Figure 7 reveals that the majority of the twentieth-century CMIP3 simulations generate a stronger tropical SST warming than NH extratropical SST warming and that the increase in meridional SST gradients is considerably greater than the increase in zonal SST gradients. The top and bottom panels of Fig. 7 further relate the simulated Mediterranean precipitation change (1971–2010 minus 1902–70) to the simulated meridional and zonal SST gradient changes, respectively. The former is the difference of tropical minus NH extratropical SST change, whereas the latter is the difference between tropical Indian Ocean minus tropical Pacific Ocean SST change. The analysis reveals a modest linear relationship; model simulations having stronger meridional and zonal increases in SST gradients tend to also generate greater Mediterranean drying.

The idealized SST forcing experiments provide independent confirmation for SST gradient–Mediterranean rainfall relationships as suggested by the empirical analyses of Fig. 7. Figure 8 (left panel) first shows the wintertime precipitation response in the SST-forced



## Sensitivity to Meridional SST Contrast



## Sensitivity to Zonal SST Contrast

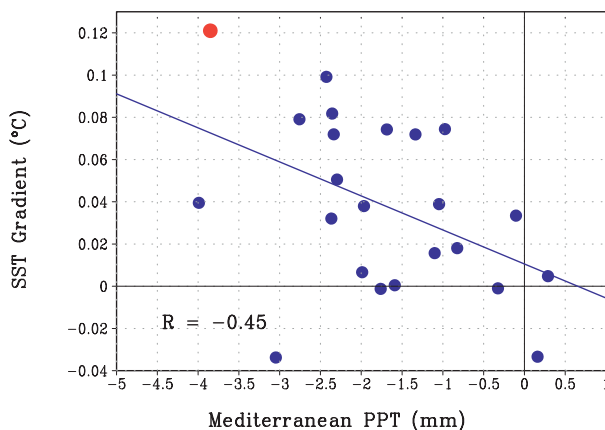


FIG. 7. Relationship between the change (1971–2010 minus 1902–70) in cold season Mediterranean precipitation (mm) and sea surface temperatures gradient ( $^{\circ}\text{C}$ ) for the 22 IPCC models: (top) area-averaged precipitation vs the meridional ( $30^{\circ}$ – $60^{\circ}\text{N}$  minus  $30^{\circ}\text{S}$ – $30^{\circ}\text{N}$ ) SST gradient and (bottom) as above but for the zonal (tropical Indian ocean minus tropical Pacific) zonal SST gradient—the observed value is indicated by the solid red dot.

AGCM experiments for uniform  $+0.5^{\circ}\text{C}$  warming of worldwide SSTs. A dry signal occurs over the eastern Mediterranean that is similar to the dry signal in that region occurring in CMIP3 and AMIP simulations. However, it is evident that such a scenario of uniform warming of  $+0.5^{\circ}\text{C}$  having no changes in SST gradients cannot be invoked to explain the Mediterranean drying that spans Gibraltar to the Middle East in nature and in experiments using realistic twentieth-century forcing.<sup>2</sup>

<sup>2</sup> Furthermore, the tropical rainfall pattern associated with the uniform  $+0.5^{\circ}\text{C}$  warming is fundamentally different from the rainfall pattern during the warm phase of ENSO (not shown).

For a scenario of warming of  $+0.5^{\circ}\text{C}$  that is restricted to tropical oceans only, a more widespread Mediterranean drying signal emerges (middle panel). A well-defined dipole pattern having wet northern Europe and dry southern Europe also emerges in response to a tropical SST warming, one bearing strong resemblance to signals in the realistically forced model suites and also to the observed change pattern (cf. Fig. 1). Also shown in Fig. 8 (right panel) is the precipitation response to a tropical SST warming that includes a zonal contrast having stronger Indian Ocean warming than the adjacent Pacific. This additional factor yields a considerable increase in the intensity of Mediterranean drying (and also northern Europe wetness), although the spatial pattern of precipitation response is similar to the response pattern to a uniform tropical ocean warming.

### b. Atmospheric circulation sensitivity

Regional atmospheric circulation over the North Atlantic exhibits strong sensitivity to spatial gradients in the overall SST change pattern (Fig. 9). Not surprisingly, when the atmospheric model is subjected to a globally uniform SST increase, the 500-hPa height signal is one of increased heights over the entire North Atlantic/European region (top left). Yet, a modest asymmetry in the response can also be discerned in which heights rise more over subtropical latitudes than over the far North Atlantic. This spatial structure is made more evident by plotting the residual height response, which is calculated by removing the global mean height anomaly from each grid point (bottom left panel). One can readily infer, from the anomalous meridional gradient in the residual field, an increase in westerly flow across most of the North Atlantic and Europe, and that such an onshore fetch would yield increased precipitation over Europe as seen in Fig. 8.

The asymmetric character of North Atlantic height responses becomes much more pronounced, and, thus, the associated westerly flow response becomes stronger when spatial gradients are added to the SST forcing. For instance, a distinct negative 500-hPa height response emerges in simulations forced with warming confined to the tropical oceans only (Fig. 9, top middle), and a well-defined anomalous anticyclone now spans much of the Mediterranean region (see Fig. 9, bottom middle) that is consistent with the expanded drying that envelops the entire Mediterranean region in these simulations. It should be noted, however, that this 500-hPa height response is materially different from a NAO structure (e.g., Barnston and Livezey 1987); its centers are southward shifted by nearly  $10^{\circ}$  latitude compared to the NAO centers of action.

When an additional spatial gradient is added to the SST forcing via enhanced Indian Ocean warmth, a stronger north–south contrast in the North Atlantic

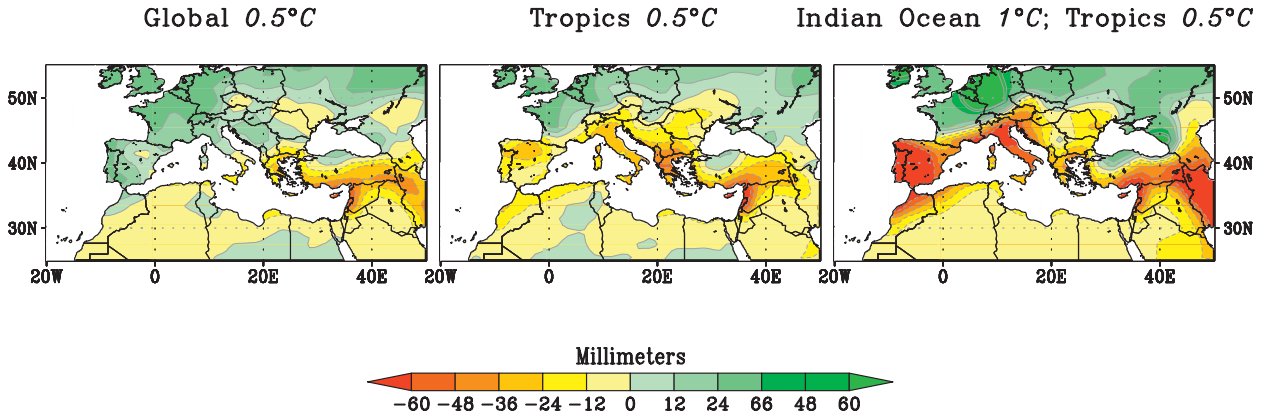


FIG. 8. Forced response of cold season (November–April) precipitation (mm) to three SST scenarios. The forced response is the average of 30 simulations using each of three scenarios of changes in SST: (left) uniform warming of +0.5°C, (middle) warming of +0.5°C in the tropics only, and (right) warming of +1°C over the Indian Ocean with +0.5°C warming elsewhere in the tropics.

height response occurs, whose centers of action are more closely aligned with the NAO centers (right panels). Low pressure anomalies occur over the Arctic while opposite-signed anomalies occur over the central North Atlantic, as typically is seen with positive index phases of the NAO, although the zonal scale of the height response is much greater than what occurs during NAO events. We also find a pronounced northward shift and an intensification of storm track anomalies in these experiments (2.5–6-day

bandpass 500-hPa height variances, not shown) that mimic the North Atlantic storm track anomalies associated with the positive NAO phase (e.g., Hurrell 1995). These dynamically driven synoptic-eddy effects are likely instrumental in explaining the substantial increase in western Mediterranean drying occurring in the warmed Indian Ocean scenario experiment compared to either the globally uniform or tropically uniform warming experiments (see Fig. 8, right).

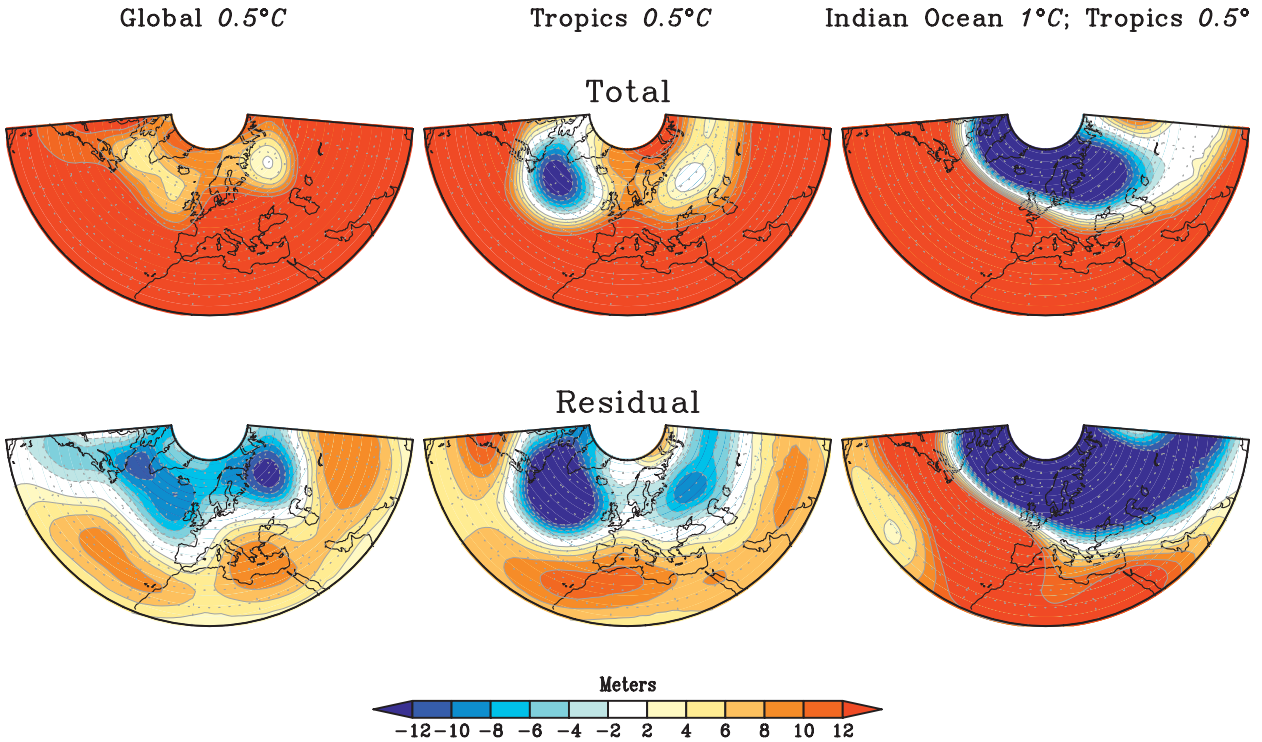


FIG. 9. As in Fig. 8 but for the forced response of 500-hPa heights (gpm) to the three SST scenarios: (top) the total height anomaly and (bottom) the residual (global mean removed) height anomaly.

Several aspects of the Mediterranean climate sensitivity revealed in these idealized SST forcing experiments are germane to understanding the recent observed drying in that area. A globally uniform  $+0.5^{\circ}\text{C}$  warming alone produces drying of the eastern Mediterranean, and this may identify a major factor responsible for the strong observed eastern Mediterranean drying. Furthermore, a zonally uniform warming of  $+0.5^{\circ}\text{C}$  of tropical SSTs alone renders widespread Mediterranean drying, and these model results may point to a major factor for the observed Mediterranean-wide drying that has taken place in concert with an observed  $+0.5^{\circ}\text{C}$  greater tropical SST warming relative to the extratropical oceans since 1902. Neither of these SST forcings excites an atmospheric response resembling the NAO pattern and thus fails to describe one aspect of the observed (based on reanalysis data) circulation change since 1902. The atmospheric responses in these simulations do, however, generate regional circulation patterns suggestive of storm track shifts conducive for Mediterranean dryness. An *isolated* and *regionally enhanced* Indian Ocean warming is thus not a sole, nor even a necessary, pattern for oceanic forcing of Mediterranean drying. The final simulation suite indicates that a SST condition that includes a stronger Indian Ocean warming relative to adjacent oceans is conducive for generating a North Atlantic/European regional circulation pattern resembling the positive polarity of the NAO, which enhances the regional drying, especially of the western Mediterranean (see Fig. 7). The modest upward trend of an index of the NAO since 1902 may thus be consistent with the influence of such remote forcing. Similarly, the lack of such an NAO trend in the CMIP simulations also appears to be physically consistent with the more uniform increase of SSTs throughout the tropics that occurs in response to external radiative forcing since 1902 in those models. Section 5 further discusses the connections among Mediterranean drying, Indian Ocean SST forcing, NAO variability, and links to anthropogenic forcing.

The above interpretation of the Mediterranean drying invokes the notion of teleconnections having a strong regional expression, but theories of anthropogenic climate change also include the expectation for a poleward expansion of the subtropical dry zones through mechanisms involving the zonal mean circulation. To what extent is the Mediterranean drying occurring in these idealized SST forcing experiments simply a response to a poleward expansion of the model's Hadley cell? The model simulations for the globally uniform warming indeed exhibit a significant poleward shift of the northern edge of the Hadley cell during November–April, which contributes to the drying of lower midlatitudes (not shown). However, there is little further Hadley cell

change in the runs that involve the addition of SST gradients (not shown), and thus the envelopment of all the Mediterranean in dryness and with greater intensity in these simulations is primarily due to mechanisms of regional teleconnection processes rather than zonally symmetric dynamics.

*c. Comparison with CMIP3, AMIP, and observations*

Principal features of CMIP3, AMIP, and observed twentieth-century tropospheric height changes can be understood in the framework of our simplified forcing experiments. Figure 10 shows the height change simulated in CMIP3 (left) and AMIP (center) and observed (right), while Table 1 presents the 500-hPa height pattern correlations (computed over the map domains of Figs. 9 and 10) of each SST scenario with the CMIP3, AMIP, and observed change patterns. For CMIP3, the relatively uniform increase in North Atlantic/European 500-hPa heights is consistent with the relatively uniform pattern of SST change simulated in response to anthropogenic greenhouse gas forcing. The modest amplitude asymmetric component of the CMIP3 residual height change pattern consists of local minima over the North Atlantic and eastern Europe that are very similar to those occurring in the globally uniform and also the tropically uniform warming runs; they are materially different, however, from those occurring in the idealized SST runs subjected to enhanced Indian Ocean warming. Table 1 indicates a 0.60 and 0.83 pattern correlation of CMIP3 with height responses in the globally uniform and tropically uniform SST warming runs, respectively, but only a 0.18 correlation with the height response in the enhanced Indian Ocean warming experiments.

In contrast, the AMIP simulations (Fig. 10, middle panels) yield a twentieth-century circulation change that projects more strongly on a NAO structure and, in this sense, is most similar to the height response occurring in the idealized SST runs using enhanced Indian Ocean warming. We note also that the zonal scale of the height change pattern in AMIP is considerably larger than what occurs during NAO events, yet appears consistent with a footprint attributable to an influence of Indian Ocean SST warming. Table 1 indicates a 0.86 pattern correlation between the AMIP height change pattern and the height response in the enhanced Indian Ocean warming runs, whereas little spatial agreement exists with the height responses in the other two idealized SST experiments. A similar assessment pertains to the observed height change pattern. The twentieth-century reanalysis data indicates a height change pattern since 1902 that also projects onto a positive NAO pattern. Table 1 confirms that this most resembles the forced response involving enhanced warming

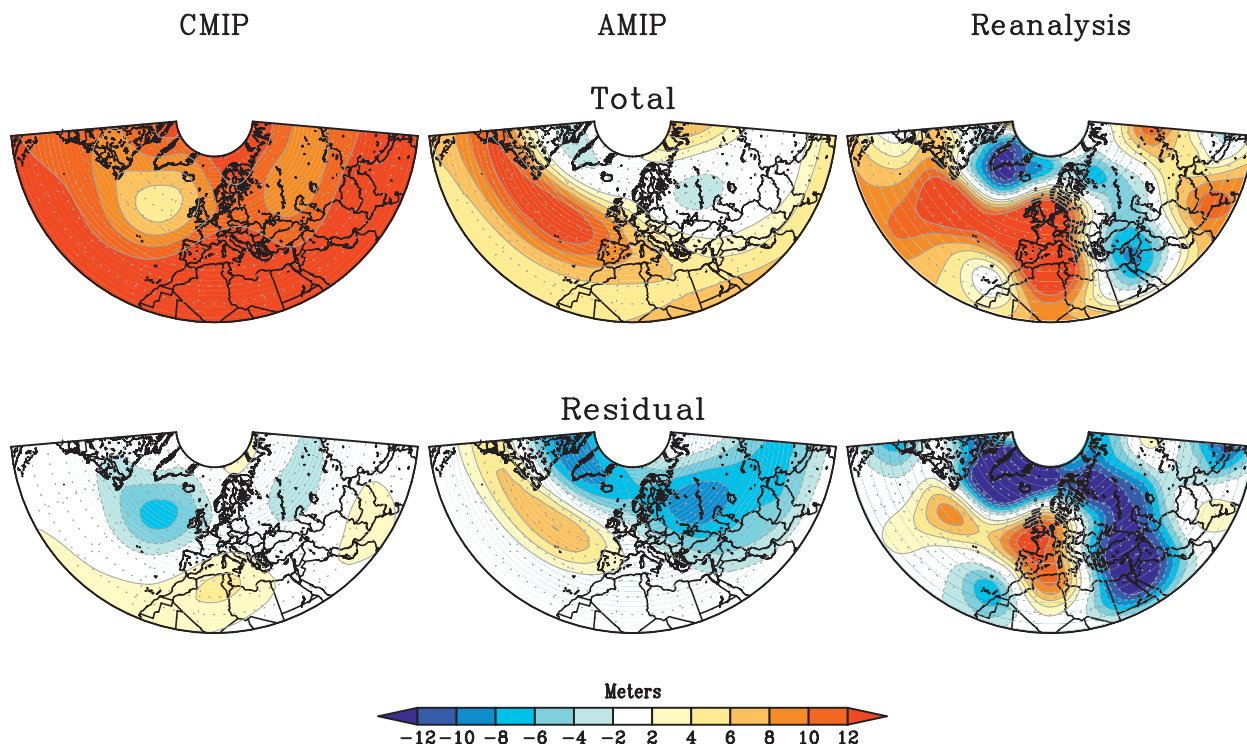


FIG. 10. Cold season (November–April) 500-hPa height departures (gpm) for the period 1971–2010 minus 1902–70 for (left) the 22-model IPCC ensemble average, (middle) the 40-run ensemble average of twentieth century AMIP runs, and (right) based on twentieth-century reanalysis: (top) total height changes and (bottom) the residual fields with global mean height anomaly removed.

of the Indian Ocean with a 0.61 pattern correlation, whereas appreciably weaker correlations of about 0.3 describe the agreement with height anomaly patterns in the other two experiments.

## 5. Conclusions

### a. Summary

Our results indicate that a change in the Mediterranean region cold-season precipitation has likely occurred over the period 1902–2010. The direction of this change has been toward drier conditions with increased drought frequency after about 1970. Our confidence language reflects uncertainties in the magnitude of the actual observed drying, and possible biases in unforced coupled model estimates of natural internal variability. The results further indicate that the detected change toward drier conditions is attributable, *in part*, to the Mediterranean region's sensitivity to time-evolving external radiative forcing. The *amplitude* of the externally forced, area-averaged Mediterranean drying signal (estimated from the ensemble mean of CMIP3 simulations) is roughly one-half the magnitude of the observed drying, indicating that other processes likely also contributed to the observed drying. Regarding mechanisms

for Mediterranean drying, our analysis of AMIP simulations reveals that SST forcing alone yields a signal of Mediterranean drying whose spatial signature and temporal evolution capture several elements of both observations and the CMIP3 externally forced signal. SST forcing has thus likely played an important part in the observed Mediterranean drying, and the externally forced drying signal likely also occurs through the impact of a SST change signal.

This line of reasoning motivated further experimentation in which atmospheric models were subjected to idealized forcings describing several gross features of global

TABLE 1. Spatial correlation of 500-hPa height anomaly patterns for the North Atlantic/European region of the map domains shown in Figs. 8 and 9. The height responses in the atmospheric model simulations using idealized SST forcings (top row) are compared to the twentieth-century height change patterns in the realistically forced model experiments and observations (left column).

	Pattern correlations for 500-hPa height responses		
	Uniform global SST warming	Uniform tropical SST warming	Enhanced Indian Ocean warming
CMIP3	0.60	0.83	0.18
AMIP	0.17	0.21	0.86
Obs	0.32	0.29	0.61



SST change during the last century. A uniform  $+0.5^{\circ}\text{C}$  warming of the global oceans induced strong drying over the eastern Mediterranean region but was ineffective in producing the distinctive pattern of a widespread Mediterranean drying trend occurring in observations and CMIP3, and AMIP simulations for 1902–2010. Instead, meridional and zonal gradients in the SST change pattern were shown to be necessary to induce a Mediterranean precipitation response having a single sign that spanned from Gibraltar to the Middle East. For  $+0.5^{\circ}\text{C}$  warming confined to tropical latitudes only (thereby introducing an enhancement of the climatological meridional gradient), drying envelops the entire Mediterranean region. A regional atmospheric circulation response in these uniform warming runs, resembling the CMIP3 simulated circulation change pattern, is of a type that, though different from a NAO pattern, is conducive for displacing moisture bearing storms northward into the European continent. The intensity of this drying increases substantially, especially over the western Mediterranean, when the tropical Indian Ocean is warmed  $+0.5^{\circ}\text{C}$  more than the rest of the tropical oceans (thereby introducing an enhancement of the climatological zonal gradient). For this latter forcing scenario, the atmospheric response pattern was shown to project onto the positive phase of the NAO. Although this additional Indian Ocean warmth exaggerates the magnitude of observed warming of the Indian Ocean relative to the Pacific Ocean during the past century, the experimental results may nonetheless point to a factor contributing to the modest upward trend in a NAO index in observations and AMIP simulations during 1902–2010 and may also account for the lack thereof in the CMIP3 ensemble mean (which produced very little zonal gradient in the tropical SST response to external radiative forcing). Of course, other processes can also induce low frequency variability of the North Atlantic atmospheric circulation, and the possibility that the modest amplitude observed upward trend toward a positive phase of the NAO was due to internal atmospheric variability alone cannot be discounted.

### *b. Discussion*

In reconstructing a history of Mediterranean climate since 1902 and in determining the causes for its change we have relied upon various sources. The description of precipitation change itself is not without its difficulties, given the rather sparse network of long-term observations, and, while four different analyses agree that a large spatial average of Mediterranean land precipitation has almost certainly declined during 1902–2010, the intensity of that decline is less certain. The description of the atmospheric circulation change at 500 hPa was based upon a reanalysis dataset that assimilated only surface pressure

reports and used monthly SST and sea ice distributions (Compo et al. 2011). The authors' assessment of that data indicates a particularly high quality of extratropical NH circulation features that is comparable to that of current three-day operational numerical weather prediction forecasts. The use of lower-boundary forcings to constrain the assimilating model implies that the long-term change estimated from the reanalysis is not strictly independent of AMIP- or CMIP-style simulations of change, though it is very likely that the use of observed sea level pressure data provides the major constraint for extratropical circulation in the reanalysis data. We thus have high confidence that the reanalysis and climate simulations are independent sources of information regarding the region's climate change since 1902 and that these simulations provide a meaningful appraisal of probable causes for the increase in Mediterranean drought frequency.

Our results indicate that similar physical processes have likely contributed to observed and CMIP3-simulated Mediterranean wintertime drying (i.e., SST forcing and atmospheric teleconnections), although some details of the SST change since 1902 differ between observations and the CMIP3 externally forced signal that impact details of the regional atmospheric response patterns. Regarding this latter issue, our results suggest that the Indian Ocean may have played a greater role in the observed Mediterranean drying than occurs in CMIP3, including the rendering of a NAO circulation response in observations that is lacking in CMIP3. The importance of positive SST anomalies over the Indian Ocean warm pool for widespread Mediterranean drought associated with a regional NAO response has been shown in several prior studies (e.g., Hoerling et al. 2001; Bader and Latif 2003; Lu et al. 2004; Hoerling et al. 2004; Li et al. 2006), and our results are consistent with those. A new feature of the simulations conducted here is that Indian Ocean warming, when immersed in a pattern of uniform warming of the world oceans or of uniform warming of the tropical oceans, is ineffective in generating a NAO-like response pattern. Thus, to the extent that the true anthropogenic signal of SST change during 1902–2010 is one of uniform warming, an NAO-like atmospheric change is not an expectation (at least based on SST forcing considerations alone), and this helps to explain the lack of a NAO signal in CMIP3. Yet, the observed SST change since 1902, contrary to the signal of SST change estimated from the ensemble of CMIP3 runs, does possess an appreciably stronger Indian Ocean warming than occurring over the tropical Pacific, a condition that our idealized SST experiments and those in prior studies indicates would favor a NAO-like response. In this regard, the AMIP simulations driven by observed SST variations simulate the observed

500-hPa height change pattern better than do the CMIP3 simulations driven by external radiative forcing. We thus conclude that observed NAO-like circulation change is reconcilable with a fingerprint of regional sensitivity to enhanced Indian Ocean warming (relative to the tropical Pacific), consistent with previous studies.

Despite this ability to explain the behavior in the fully forced time-varying climate simulations, some uncertainty must necessarily remain regarding a full understanding of how, and by how much, anthropogenic forcing contributed to *observed* Mediterranean drying during 1902–2010. It is especially unclear to what extent the observed zonal contrast in tropical SST warming, which contributes to drying via an NAO teleconnection link, is due to internal variability or is an expression of nature's response of anthropogenic forcing. There is no evidence that indicates the anthropogenic signal of the tropical ocean response is one of stronger Indian Ocean warming relative to the Pacific Ocean. It is plausible that observational indications for an enhanced zonal gradient is linked to a weaker warming in the tropical Pacific resulting from a particularly strong occurrence of internal variability in that region (rather than an enhanced anthropogenic warming signal over the Indian Ocean), as suggested by results of Knutson et al. (2006) and more recently by Hoerling et al. (2010). On the other hand, it is also plausible that CMIP3 models have biases in their SST response to GHG forcing, especially over the tropical Pacific (e.g., Vecchi et al. 2008) and that the true tropical ocean response deviates appreciably from the ensemble mean of CMIP3 simulations. This line of reasoning would argue that an anthropogenic drying signal of the Mediterranean region is stronger than inferred from the CMIP3 ensemble mean simulations.

A key point that our results raise, however, is that the link between anthropogenic forcing and the observed Mediterranean drying during the analysis period of 1902–2010 does not require a connection involving *enhanced* Indian Ocean warming nor a North Atlantic Oscillation circulation response—a uniform warming of the tropical oceans alone induces widespread drying. The stronger observed drying occurring since 1902 (than in the CMIP3 ensemble) may result from an internally generated component of SST variability over the tropical Pacific, where a minimum of warming has acted to intensify the tropical-wide SST contrast in such a manner that favors forcing of a trend toward the positive NAO phase. The stronger observed drying may also merely reflect the serendipitous contribution by internal atmospheric variability. Each of these internal factors, or a combination of them, could have enhanced the Mediterranean drying in observations above and beyond the anthropogenic signal derived from the CMIP3 ensemble mean.

Uncertainties in the SST analyses themselves, however, do not permit a more definitive assessment of the difference between observed and CMIP3 SST gradients across the Indo-Pacific region. Various historical analyses lead to differences in the magnitudes of zonal gradients in SST (Deser et al. 2010). It is nonetheless reasonable to believe that an appreciable west–east gradient in tropical SST change exists in light of the agreement between AMIP and observed circulation changes, in the form of a positive NAO structure. This agreement may be a proxy indicator for a greater Indian Ocean warming compared to the tropical Pacific warming, as implied also by the results of our idealized SST sensitivity experiments.

There are further questions concerning Mediterranean drought to which our analysis applies, though which cannot be answered based on the current results alone. In particular, what is the actual intensity of the anthropogenic drying signal over the Mediterranean region? The observed drying has been roughly double the ensemble mean CMIP3 drying, and only one of the 22 models generate a drying as strong as observed. We have already addressed the fact that the true amplitude of the observed drying is not known, and thus the difference with CMIP3 amplitudes cannot be absolutely determined. It is nonetheless plausible that the observed change may include a substantial internal variability component that is coincidentally of the same polarity as the climate change signal. Indeed, the spread of PDFs in Fig. 4 attests to the considerable amplitude of internally generated centennial-scale variations of Mediterranean precipitation trends. It is also possible that the true climate change signal is stronger than implied by the CMIP3 ensemble mean. For instance, if the climate change signal of tropical SST change involves a stronger zonal SST gradient across the Indo-Pacific than simulated in CMIP3, but akin to observations, an intensified Mediterranean drying signal would be expected. This line of reasoning would argue that the amplitude of the CMIP3 drying signal is erroneously weak owing to biases in the coupled model's global SST response. Yet, our AMIP results also generate a drying signal that is only about one-half the amplitude of the observed drying, and they also contain only one simulation (of 40) that has a drying trend greater than observed. Further climate simulations studies are required to clarify these matters. It is expected that climate models to be used in support of the next IPCC Assessment, for instance, will use considerably higher spatial resolution than those used in CMIP3 and in our study. It will be useful to repeat the current analysis with them and especially to conduct parallel large ensemble suites of CMIP- and AMIP-style integrations that use the same atmospheric models.

A second question concerns the subregional precipitation signals of climate change. We have already indicated that the observational data are not adequate to assess centennial changes at local scales. There are nonetheless tantalizing indications of particularly strong drying over portions of southern Spain, Morocco, northern Algeria, and portions of the Middle East where select stations are available (see Fig. 2). These are among the areas that we have found to be especially drought sensitive when Indian Ocean warming is enhanced, raising the scientific need to further understand how climate change will affect the detailed pattern of SST change. The question is of considerable practical importance also in light of assessments for current and projected 2025 water scarcity in the Mediterranean (Mediterranean Water Scarcity and Drought Working Group 2007). It is estimated that water withdrawals—unrelated to climate change—will alone exceed the average annual volume of renewable natural water resources in Spain's Mediterranean basin, North Africa, and the eastern Mediterranean by 2025. A climate change-induced drying will clearly aggravate a significant water resource challenge for the area, and information on the subregional patterns of climate change will be important in assisting a planning framework for societal responses to drought.

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#### REFERENCES

- Alley, R. B., 2007: Summary for policymakers. *Climate Change 2007: The Physical Science Basis*, S. Solomon et al. Eds., Cambridge University Press, 1–18.
- Bader, J., and M. Latif, 2003: The impact of decadal-scale Indian Ocean SST anomalies on Sahelian rainfall and the North Atlantic Oscillation. *Geophys. Res. Lett.*, **30**, 2169, doi:10.1029/2003GL018426.
- Barnston, A., and R. Livezey, 1987: Classification, seasonality, and persistence of low-frequency atmospheric circulation patterns. *Mon. Wea. Rev.*, **115**, 1083–1126.
- Compo, G. P., and Coauthors, 2011: The Twentieth Century Reanalysis Project. *Quart. J. Roy. Meteor. Soc.*, **137**, 1–28, doi:10.1002/qj.776.
- Davis, S. M., and K. H. Rosenlof, 2012: A multidagnostic intercomparison of tropical-width time series using reanalyses and satellite observations. *J. Climate*, **25**, 1061–1078.
- Delworth, T. L., and Coauthors, 2006: GFDL's CM2 global coupled climate models. Part I: Formulation and simulation characteristics. *J. Climate*, **19**, 643–674.
- Deser, C., A. Phillips, and M. Alexander, 2010: Twentieth-century tropical sea surface temperature trends revisited. *Geophys. Res. Lett.*, **37**, L10701, doi:10.1029/2010GL043321.
- ESPERE, cited 2006: Climate encyclopedia—Food and climate. [Available online at <http://espere.mpch-mainz.mpg.de/documents/pdf/>.]
- Gillett, N. P., 2003: Climate change and the North Atlantic Oscillation. *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*, *Geophys. Monogr.*, Vol. 134, Amer. Geophys. Union, 193–209.
- Giorgi, F., 2006: Climate change hot spots. *Geophys. Res. Lett.*, **33**, L08707, doi:10.1029/2006GL025734.
- Hoerling, M. P., J. W. Hurrell, and T. Xu, 2001: Tropical origins for recent North Atlantic climate change. *Science*, **292**, 90–92.
- , —, —, G. T. Bates, and A. Phillips, 2004: Twentieth-century North Atlantic climate change. Part II: Understanding the effect of Indian Ocean warming. *Climate Dyn.*, **23**, 391–405.
- , J. Eischeid, and J. Perlwitz, 2010: Regional precipitation trends: Distinguishing natural variability from anthropogenic forcing. *J. Climate*, **23**, 2131–2145.
- Hulme, M., T. J. Osborn, and T. C. Johns, 1998: Precipitation sensitivity to global warming: Comparison of observations with HadCM2 simulations. *Geophys. Res. Lett.*, **25**, 3379–3382.
- Hurrell, J. W., 1995: Decadal trends in the North Atlantic Oscillation: Regional temperatures and precipitation. *Science*, **269**, 676–679.
- , and C. Deser, 2009: North Atlantic climate variability: The role of the North Atlantic Oscillation. *J. Mar. Syst.*, **78**, 28–41, doi:10.1016/j.jmarsys.2008.11.026.
- , M. P. Hoerling, A. Phillips, and T. Xu, 2004: Twentieth-century North Atlantic climate change. Part I: Assessing determinism. *Climate Dyn.*, **23**, 375–390.
- Jones, P., T. Osborn, and K. Briffa, 2003: Pressure-based measures of the North Atlantic Oscillation (NAO): A comparison and an assessment of changes in the strength of the NAO and its influence on surface climate parameters. *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*, *Geophys. Monogr.*, Vol. 134, Amer. Geophys. Union, 51–62.
- Kiehl, J., J. Hack, G. Bonan, B. Boville, D. Williamson, and P. Rasch, 1998: The National Center for Atmospheric Research Community Climate Model: CCM3. *J. Climate*, **11**, 1131–1149.
- Knutson, T., and Coauthors, 2006: Assessment of twentieth-century regional surface temperature trends using the GFDL CM2 coupled models. *J. Climate*, **19**, 1624–1651.
- Legates, D. R., and C. J. Willmott, 1990: Mean seasonal and spatial variability global surface air temperature. *Theor. Appl. Climatol.*, **41**, 11–21.
- Li, S., M. Hoerling, and S. Peng, 2006: Coupled ocean–atmosphere response to Indian Ocean warmth. *Geophys. Res. Lett.*, **33**, L07713, doi:10.1029/2005GL025558.

- Lu, J., R. J. Greatbatch, and K. A. Peterson, 2004: Trend in Northern Hemisphere winter atmospheric circulation during the last half of the twentieth century. *J. Climate*, **17**, 3745–3760.
- , G. Vecchi, and T. Reichler, 2007: Expansion of the Hadley cell under global warming. *Geophys. Res. Lett.*, **34**, L06805, doi:10.1029/2006GL028443.
- Mariotti, A., 2010: Recent changes in the Mediterranean water cycle: A pathway toward long-term regional hydroclimatic change? *J. Climate*, **23**, 1513–1525.
- , and A. Dell'Aquila, 2011: Decadal climate variability in the Mediterranean region: Roles of large-scale forcings and regional processes. *Climate Dyn.*, doi:10.1007/s00382-011-1056-7, in press.
- , N. Zeng, J.-H. Yoon, V. Artale, A. Navarra, P. Alpert, and L. Z. X. Li, 2008: Mediterranean water cycle changes: Transition to drier 21st century conditions in observations and CMIP3 simulations. *Environ. Res. Lett.*, **3**, 044001, doi:10.1088/1748-9326/3/4/044001.
- Mediterranean Water Scarcity and Drought Working Group, 2007: Technical report on water scarcity and drought management in the Mediterranean and the water framework directive. Mediterranean EU Water Initiative/Water Framework Directive Joint Process Tech. Rep. 009-2007, 133 pp.
- Meehl, G., and Coauthors, 2007: The WCRP CMIP3 multimodel dataset: A new era in climate change research. *Bull. Amer. Meteor. Soc.*, **88**, 1384–1394.
- Mehta, V., M. Suarez, J. Manganello, and T. Delworth, 2000: Oceanic influence on the North Atlantic Oscillation and associated Northern Hemisphere climate variations: 1959–1993. *Geophys. Res. Lett.*, **27**, 121–124.
- Nelson, G. C., and Coauthors, 2009: Climate change: Impact on agriculture and costs of adaptation. IFPRI Food Policy Rep., 19 pp. [Available online at doi:10.2499/0896295354.]
- Osborn, T., 2011: Variability and changes in the North Atlantic Oscillation index. *Hydrological, Socioeconomic and Ecological Impacts of the North Atlantic Oscillation in the Mediterranean Region*, S. M. Vicente-Serrano and R. M. Trigo, Eds., Advances in Global Change Research, Vol. 46, Springer, 9–22.
- Previdi, M., and B. G. Liepert, 2007: Annular modes and Hadley cell expansion under global warming. *Geophys. Res. Lett.*, **34**, L22701, doi:10.1029/2007GL031243.
- Robertson, A., C. Mechoso, and Y. Kim, 2000: The influence of Atlantic sea surface temperature anomalies on the North Atlantic Oscillation. *J. Climate*, **13**, 122–138.
- Rudolf, B., and U. Schneider, 2005: Calculation of gridded precipitation data for the global land-surface using in-situ gauge observations. *Proc. Second Workshop of the Int. Precipitation Working Group (IPWG)*, Monterey, CA, IPWG, 231–247.
- Schubert, S. D., M. J. Suarez, P. J. Pegion, R. D. Koster, and J. T. Bacmeister, 2004: On the cause of the 1930s Dust Bowl. *Science*, **33**, 1855–1859.
- Seager, R., and Coauthors, 2007: Model projections of an imminent transition to a more arid climate in southwestern North America. *Science*, **316**, 1181–1184.
- , N. Naik, and G. A. Vecchi, 2010: Thermodynamic and dynamic mechanisms for large-scale changes in the hydrological cycle in response to global warming. *J. Climate*, **23**, 4651–4668.
- Smith, T. M., and R. W. Reynolds, 2005: A global merged land, air, and sea surface temperature reconstruction based on historical observations (1880–1997). *J. Climate*, **18**, 2021–2036.
- Solomon, S., D. Qin, M. Manning, M. Marquis, K. B. Averyt, M. Tignor, H. L. Miller Jr., and Z. Chen, Eds., 2007: *Climate Change 2007: The Physical Science Basis*. Cambridge University Press, 996 pp.
- Sutton R., W. Norton, and S. Jewson, 2000: The North Atlantic Oscillation—What role for the ocean? *Atmos. Sci. Lett.*, **1**, 89–100, doi:10.1006/asle.2000.0018.
- Ulbrich, U., and Coauthors, 2006: The Mediterranean climate change under global warming. *Mediterranean Climate Variability and Predictability*, P. Lionello et al., Eds., Elsevier, 398–415.
- Vecchi, G. A., A. Clement, and B. J. Soden, 2008: Pacific signature of global warming: El Niño or La Niña? *Eos, Trans. Amer. Geophys. Union*, **89**, doi:10.1029/2008EO090002.
- Vose, R. S., R. L. Schmoyer, P. M. Steurer, T. C. Peterson, R. Heim, T. R. Karl, and J. K. Eischeid, 1992: The Global Historical Climatology Network: Long-term monthly temperature, precipitation, sea-level pressure, and station pressure data. Carbon Dioxide Information Analysis Center Rep. ORNL/CDIAC-53, 25 pp. [Available from Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, TN 37831-6335.]
- Wunsch, C., 1999: The interpretation of short climate records, with comments on the North Atlantic and Southern Oscillations. *Bull. Amer. Meteor. Soc.*, **80**, 245–256.
- Yin, J. H., 2005: A consistent poleward shift of the storm tracks in simulations of 21st century climate. *Geophys. Res. Lett.*, **32**, L18701, doi:10.1029/2005GL023684.
- Zhang, X., F. W. Zwiers, G. C. Hegerl, F. H. Lambert, N. P. Gillett, S. Solomon, P. A. Stott, and T. Nozawa, 2007: Detection of human influence on twentieth-century precipitation trends. *Nature*, **448**, 461–465.