USING AN IMPROVED COUPLING DROUGHT INDEX TO PREDICT THE PERSISTENCE OF METEOROLOGICAL DROUGHT OVER WEST CENTRAL INDIA

by

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LIST OF ABBREVIATIONS AND SYMBOLS

2m temperature: air temperature at 2m above the surface

ABL: atmospheric boundary layer

AD: Anderson-Darling

CDF: cumulative distributive function

CDI: coupling drought index

CDI_{imp}: improved coupling drought index

CFSv2: NCEP Climate Forecast System Reanalysis version 2

CMAP: CPC Merged Analysis of Precipitation

c_p: specific heat of air at constant pressure

CPC: Climate Prediction Center

CSFR: NCEP Climate Forecast System Reanalysis

CTP: convective triggering potential

DALR: dry adiabatic lapse rate

e: vapor pressure

ECMWF: European Centre for Medium-Range Weather Forecasts

EDA: ensemble of daily assimilations

ELR: environmental lapse rate (γ)

ENSO: El Nino/Southern Oscillation

ERA5: ECMWF climate reanalysis version 5

g: acceleration due to gravity (grav)

ɤ: psychometric constant (e_p)

HI: humidity index

hPa: hectapascal(s)

IMD: Indian Meteorological Department

IOD: Indian Ocean Dipole

JJAS: June, July, August, and September

J: joule(s)

K: Kelvin

kg: kilogram(s)

KS: Kolmogorov-Smirnov

MALR: Moist Adiabatic Lapse Rate

MAM: March, April, and May

mb: millibars

MERRA2: Modern-Era Retrospective Analysis for Research and Applications version 2

mid: middle

MLE: maximum likelihood estimator

NCEP: National Centers for Environmental Prediction

N_d: number of dry coupling days

N_t: total number of days

N_w: number of wet coupling days

OND: October, November, and December

Pa: Pascal(s)

par: parcel

q: specific humidity

q 2m: specific humidity at 2m above the surface

R_d: molar constant of dry air

seg: segment

SM: soil moisture

SON: September, October and November

SPI: standardized precipitation index

T: temperature

T_d: dewpoint temperature

T_{env}: temperature of the observed profile

Tv: virtual temperature

z: atmospheric height

α: significance level

λ: latent heat of vaporization

ABSTRACT

USING AN IMPROVED COUPLING DROUGHT INDEX TO PREDICT THE PERSISTENCE OF METEOROLOGICAL DROUGHT OVER WEST CENTRAL INDIA

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The objective of this study is to investigate the role of the atmospheric boundary layer (ABL) and hence, roles of evaporation and convection in the process of intensification or persistence of meteorological droughts in west central India. The reason of choice of this region is that it is seen to be statistically significant in terms of spatial extent of droughts from 1951-2010. Although local geographic factors influence rainfall in this region, the main source of precipitation is the Southwest monsoons during the months of June, July, August and September (JJAS); which are affected by the tropical sea surface anomalies; and hence by El Nino/Southern Oscillation (ENSO), Indian Ocean Dipole (IOD) and other synoptic systems (like monsoon depressions) which are due to the dynamical instabilities of the mean circulation. These affect the slowly changing variable soil moisture; the soil moisture anomalies then affect the summer rainfall. However, most of the land-atmosphere coupling studies conducted earlier over India ignore the role of the ABL which controls the major thermodynamics and dynamical circulations. Hence, here we try to apply the concept of coupling drought index (CDI) during the periods of monsoon onset (March, April, and May [MAM]) over the entire spatial extent over the time-period of 1979-2010; when the atmosphere represents somewhat transitional coupling conditions, to identify regions of dry and wet coupling regions. Here, however, while testing the normality of the soil moisture anomaly curves, we have replaced the Kolmogorov-Smirnov Test with the Anderson-Darling test to build an improved CDI to gain more accuracy and sensitivity. Thereby, it is expected to predict whether drought will persist, intensify, or recover over the study region. The results from the developed index indicated that the drought conditions would neither intensify nor recover over the study region for the year 2011 and would continue to persist as it was for the past three decades. The observations validate this result.

Keywords: Improved coupling drought index, meteorological drought, west central India, Anderson-Darling Test

Chapter 1: INTRODUCTION

It had been pointed out by Eagleson (1991) that the annual global mean evaporation from land surfaces is about 60% of the annual global mean precipitation. The percentage is even greater in tropics and during local summers. This shows us that evaporation from land surfaces is a very important contributing factor to the global water budget and the hydrological cycle. However, evaporation does not directly cause precipitation; as depending upon the geographic location, the evaporated moisture can get advected away from its origin. So, I need to know about the available and the precipitable moisture over an area. It has been seen that soil moisture plays an important role regarding this. Evaporation helps to regulate heating of the ground, which cools down once evaporation occurs; therefore, it influences sensible heat flux. Thus, it helps in partitioning the incoming solar energy into sensible and latent heat flux. Soil moisture also determines the rate of evaporation and, therefore, the evaporative fraction and moisture supply. All these have different values for different contents of soil moisture; i.e., dry soil, wet soil, etc. Again, these effects are not exactly locally bound. For example, when there is enough heating over the ground, that is, sensible heat flux is high, a heat low is formed which triggers dynamical circulation. The circulation may lead to intensification or dissipation of the system depending upon the nature of the circulation. The circulation may also generate another air circulation system or heat transfer pathway. Whether precipitation will occur is largely influenced by the dry or moist adiabatic lapse rate which in turn depends on the nature of dynamical circulation, its associated flow pattern, and the soil moisture memory.

The positive or negative soil moisture anomaly conditions often caused due to an atypically heavy rainfall or an unusual dry spell; take weeks or months to dissipate through evaporation and other processes to reach the mean value again. But, in effect, the soil can ‘‘remember’’ the wet or dry conditions that caused an anomaly long after these conditions have been forgotten by the atmosphere (Koster, 2001). This soil moisture persistence is known as its “memory”. Quantifying it requires multidecadal records of soil moisture. It is difficult, for such records do not exist in many parts of the world; especially over India.

We need to remember that the soil moisture is a slowly changing variable with a memory of about 90 days (Vinnikov and Yeserkepova,1991; Carson and Sangster, 1981; Rind, 1982; Delworth and Manabe, 1988). Many studies (Ek and Mahrt,1994; Eltahir and Pal,1996; Guo et al, 2006; Koster et al, 2006; Dirmeyer et al, 2006; Betts, 2009; Wei et al, 2010a, Santanello et al, 2011) over the years have confirmed that initial soil wetness can help us predict precipitation to seasonal and sub-seasonal scales.

Now, drought is generally defined in terms of the degree of dryness in comparison to some normal or average value and the duration of the dry period. As per the categorization by Wilhite and Glantz (1985), droughts can be meteorological, hydrological, agricultural and socioeconomic. Here, I deal with meteorological droughts alone which are considered region specific; and for which the atmospheric conditions that result in deficiencies in precipitation vary highly from region to region. Generally, it is characterized by precipitation deficiency in terms of amount, intensity or threshold periods of rainfall which result in reduced infiltration, runoff, deep percolation, and ground water recharge. It can also be identified with associated high temperatures, high winds, low relative humidity, greater sunshine, and less cloud cover all of which result in increased evaporation and transpiration. As discussed above, all of these lead to soil water deficiency making land-atmosphere coupling important. Over India, drought conditions are said to prevail when actual seasonal rainfall is deficient by more than twice the mean deviation (Ramdas, 1960).

For decades there were studies trying to establish a relationship between soil moisture and drought (Namias, 1989; Atlas et al, 1993; Fennessy and Shukla, 1999; Sud et al; 2003; Roundy, Ferguson & Wood, 2013; Roundy and Santanello; 2016). The surface soil moisture is representative of the uppermost 0-2cm layer. But, in some cases, positive feedback was found; whilst in others there was a clear negative feedback. But, to understand the land surface, it is important to understand the properties of the land surface and the atmospheric boundary layer above it which controls its interactions (dynamical circulations) and atmospheric thermodynamics (fluxes from the land, triggering of convection). Findell and Eltahir (2002) investigated atmospheric controls on soil moisture-boundary layer interactions and came up with a framework involving the convective triggering potential (CTP) and the humidity index (HI).

Development of the CTP and the HI are based on the premise that specific atmospheric conditions favour triggering of rainfall over wet (positive feedback) and dry (negative feedback) land surface states. CTP was derived to diagnose the crucial layer of the atmosphere in terms of the boundary layer growth. It was calculated by integrating (on a daily basis) the region between the atmospheric profile and the moist adiabatic lapse rate (MALR). The HI is a saturation deficit (dewpoint- depression).

Based on these two variables any region was divided up into four strict coupling regimes: dry soil advantage, wet soil advantage, transitional, and atmospherically controlled. In the atmospherically controlled regime, the land surface does not play much role in triggering convection. In the dry soil advantage, the atmosphere is drier, thus, having a higher HI and the atmosphere profile is closer to the dry adiabatic lapse rate. Wherever large boundary layers are formed due to high sensible heat fluxes, generating large values of CTP, convection is triggered. For wet advantage cases, the atmospheric state is closer to the wet adiabatic lapse rate. The HI values are generally low. Convection is triggered by a strong increase in the moist static energy from the land surface. The area between the dry and wet soil advantage regime, where the land surface plays a role, is known as the transitional regime. In the absence of convectional precipitation, for the dry and wet soil advantage, there is a persistence of the land surface states which again indicates that evaporation plays an integral part in the land-atmosphere coupling especially in the times of no precipitation or meteorological droughts.

Meehl (1994 a,b) modified surface conditions and found a correlation between a stronger Indian summer monsoon and the land-sea temperature contrast; and between wet soil conditions and Indian summer monsoon precipitation; a kind of positive feedback. Douville (2001) observed that rainfall increases for wetter surfaces in the case of north India. But, when land-sea contrast decreases, i.e., as the study area moves more inland away from the seas; increased water recycling is balanced by a decreased moisture convergence.

Since India was identified as one of the hotspot coupling zones by Koster et al. (2006), and since recent concurrent droughts over India have imposed serious climatological concerns, I chose to investigate this subject. I shall limit the study area to west central India where the spatial extent of droughts from 1951 to 2010 are statistically significant. This region is landlocked, where the land-sea contrast is comparatively low. It is known that India gets its rainfall from the southwest monsoons and monsoon droughts over India are affected by tropical sea surface anomalies, i.e., by ENSO. There are also additional local effects which show decreased sea breeze as a result of irrigation for south West Bengal due to reduced low-level moisture (Lohar and Pal, 1995). This shows us that precipitation is likely sensitive to large-scale irrigation over India and other places. This also puts forward the importance of land-atmosphere coupling and SM influences over precipitation.

Over India, although most works have been done investigating the droughts, very few have been done involving the role of the atmosphere boundary layer in the process. During monsoon onset and retreat, India resembles a transition zone. It is easier to study it in those periods and establish feedbacks during those times. Thus, I intend to study the monsoon onset period, marked by the months March, April and May (MAM) for 1979-2010 over west central India. The monsoon retreat period, marked by the months October, November and December (OND) (sometimes by September, October and November) are not included in this study mainly for two reasons. First, as per the Indian Meteorological Department official information, the normal date of withdrawal of the South West Monsoons from the region of study is on the 15th of October. Although the study region is affected by meteorological drought, any kind of precipitation in the region would be linked to the monsoons, and so; it has higher chances of having a positive feedback than being transitional. All other weather parameters would also show similar behavior. However, anomalous precipitation in the MAM period would only occur due to convection. Second, although Dirmeyer et al. (2009) showed that the soil moisture memory over India was highest in SON (about 20 days more); over the study area; the soil moisture memory was found to be more in MAM than SON (about 45 days more).

Namias (1959, 1960) while studying the problem of soil wetness as a boundary forcing for the atmosphere, noticed a tendency of persistence of anomalous temperatures from spring to summer as well as the tendency of wet/dry springs to precede cool/warm summers and cool/warm springs to precede wet/dry summers.

From these, he had concluded that, “moist soil may serve as a cooling reservoir by using for vaporization some of the heat normally associated with the spring to summer building of the upper level anticyclone...” (1959) and “desiccating warm and dry weather over the Plains in spring provides a healthy environment for the lodgement of the upper level anticyclone in the following summer..”(1960). Similar studies were taken forward by Rind (1982) and Karl (1983). Rind (1982) concluded that low springtime soil moisture can be looked upon as a precursor to a hot dry summer. Karl (1983) studied monthly averaged temperature and precipitation values for the United States from 1895-1981 to show similar correlations.

Oglesby and Erickson (1989) stated that a reduction in soil moisture led to increased surface temperature, with the lower atmosphere getting heated and causing low level moisture advection, one of the main factors in the persistence of droughts. Meehl (1984) while studying the role of soil moisture on monsoon rainfall over India; stated that this moisture advection becomes pivotal over inland regions; where it causes positive feedback. In such areas, even precipitation does not cause much change; and the main factor remains the increased surface heating which causes the moisture convergence in the atmosphere. So, for a chosen study area which is located inland in the Indian subcontinent, the pre-monsoon season, marked by MAM; which has very little rainfall but considerably high temperatures, is much more ideal than the post-monsoon season OND.

I intend to apply the CTP-HI feedback to verify the applicability of the mechanism over the study area and then investigate the reasons for drought with a few modifications made to the framework. If it is a positive feedback, there should be precipitation on wet soil, and no precipitation on dry soil. For a negative feedback, the reverse happens. If it is atmospherically controlled, there should be no difference between wet and dry soils.

Building on the CTP-HI framework, Roundy, Ferguson and Wood (2012) have built a coupling drought index (CDI), which is robust enough to be applied for different data sets and different regions but specific enough for use in analyzing the temporal variability of the coupling. Here, a joint probability space of CTP-HI and SM are taken at daily time steps for a chosen spatial region over a specific time period. The concept of SM anomaly is applied to this joint probability space by using the Kolmogorov-Smirnov (KS) two sample statistical test between a subspace marginal distribution and the full distribution of soil moisture. The CTP-HI-SM joint probability space is then reduced to 2D CTP-HI space, with (n x n) bins, where each bin consists of the marginal distribution of the SM and n signifies the number of bins or grids. Thus, (n x n) signifies the bin size or the grid size in the 2D space. Then, classification follows at each bin by applying the KS test at a given significance level (α). Based on these results; the bin is classified in one of the four coupling regions. It is called a dry, wet and transitional coupling if the SM in the particular bin is less than, greater than, or equal to, respectively, the value of the mean SM of the entire region. Atmospherically controlled cases are few. The process is repeated for different bins and significance levels of the KS test. Finally, CDI is defined as; CDI = (N_d - N_w)/N_t where, N_d is the number of dry coupling days, N_w is the number of wet coupling days and N_t is the total number of days during the time period. A negative value near -1 indicates consistent wet coupling or drought recovery; while a positive value near +1 indicates dry coupling or drought intensification or persistence. However, the CDI requires a vast amount of data especially from radiosondes and satellites.

The question might arise why this modification is done. While trying to test the applicability of the original CDI developed by Roundy et al. (2013) over west central India, it was found that the p-value for the KS test used in building the CDI was often too large for almost all bins and at every significance level. This means that the null hypothesis has to be rejected for the area under study and for the chosen period. Hence, the decision was taken to build an index using the Anderson-Darling test, which is a modification of the KS test but is more sensitive and thus can help in identifying droughts.

A previous study over India using a slab model by Tuinenburg et al. (2010) has confirmed the relevance of the CTP-HI framework during various seasons and provided certain numerical values for the CTP-HI as well. The slab model developed by Kim and Entekhabi (1998) assumes perfect mixing of the ABL, a cloud free ABL, no change in overlying air masses during the simulation, and constant soil moisture during the simulation.

Chapter 2: HYPOTHESIS

An improved coupling drought index can be developed to increase the sensitivity and accuracy of the test in the CTP-HI-SM framework in west central India to predict the persistence or recovery of droughts. The CDI developed by Roundy (2012) is further modified by replacing the KS statistical test with the Anderson-Darling test for fitting critical values.

It has been found that the atmospheric layer between 950 hPa and 700 hPa is critical in triggering convection. The two main indicators of stability and humidity, convective triggering potential (CTP) and the humidity index (HI) in the lower level of the atmosphere, were developed into a framework to diagnose the land-surface (i.e., SM) influence. Based on two atmospheric predictors, this framework predicts the extent of the impact of SM on convective precipitation. Based on soil moisture conditions and forcing, four kinds of cases are noted: positive feedback or wet forcing, negative feedback or dry forcing, transitional, and atmospherically controlled cases.

The CTP is a measure of stability in the lowest 1-3 km region in the ABL and is calculated by integrating the region between the atmospheric profile and the moist adiabatic temperature lapse rate.

$$CTP = \int_{P_{surf-100hPa}}^{p_{surf-300hPa}} g\left(\frac{Tv_{parcel}-Tv_{env}}{Tv_{env}}\right) dz; \quad \ldots\ldots\ldots\ldots\ldots\text{(Equation 1a)}$$

where; Tv_{parcel} is the virtual temperature of an air parcel lifted moist adiabatically from the level 100 hPa to the surface and T_{env} is the temperature of the observed profile.

CTP thus measures the buoyancy of rising air due to its difference in temperature from its surrounding environment. The greater the value, the larger the instability. When the ABL reaches to the 100 to 300 hPa level above the surface, the available energy can initiate the process of convection. The unit of CTP is joules per unit kilogram (J/kg).

The HI is a measure of the wetness of the atmosphere and is defined by the sum of the dewpoint depression at 50 and 150 hPa above the ground level. A lower value of HI indicates a more humid atmospheric state. HI_{low} is based on the index designed by Lytinska et al. (1976) whose purpose was to determine the possibility of rain for an atmospheric profile. Findell and Eltahir (2003) defined HI_{low} as;

$$HI_{low} = (T_{950} - T_{d,950}) + (T_{850} - T_{d,850}); \ldots\ldots\ldots\ldots\ldots\ldots\ldots \text{(Equation 2)}$$

where T_{950} and T_{850} are the temperatures and $T_{d,950}$ and $T_{d,850}$ are the dewpoint temperatures at 950hPa and 850hPa respectively. So, the unit of HI_{low} is K.

Though the Findell-Eltahir classification is appropriate in capturing the exchange of water from land to the atmosphere, it fails to consider the reverse scenario. The evaporative demand in the atmosphere between the dry soil advantage and the wet soil advantage regime is different because of larger (drier) values of HI (Roundy et al., 2012). So, when there is an absence of convective precipitation, evaporation plays an important role in the coupling of land-atmosphere states (Betts, 2004). Thus, in such cases, evaporation leads to the persistence of SM states. This factor of persistence becomes crucial in the context of the study of droughts. Herein, the SM anomaly is studied along with the atmospheric variables CTP and HI that trigger precipitation and evaporation to check persistence or the lack of it in the distribution of data.

For the Indian subcontinent, earlier works by Koster (2006) have shown that there is a high sensitivity between the temperature and precipitation to the land surface state, making it a “hotspot”. Further work by Guo et al (2006) has revealed that all the hotspots of coupling were located within transition zones between dry and wet climates. Between these two extremes, typical variation of evaporation is large enough to influence precipitation; but the magnitude still depends on SM. Correlations between SM state, evaporation, SM memory, and moisture recycling ratios are highest during monsoon onset (March-April-May [MAM]) for the study region concerned. This period in the year also resemble the transition zone between the wet and dry climates (Dirmeyer et al, 2009).

For the Indian subcontinent, the CTP-HI_{low} threshold values varied year to year and from station to station for the transition climate zone. So, I attempted to build an accurate coupling scheme which can analyze the temporal aspect of the land-atmosphere coupling. To do so, I follow the basic methodology followed of Roundy et al. (2012). I start by taking an average of all the SM values over space and time and then introducing the concept of SM anomaly to the joint CTP-HI-SM space, which is specific to a particular period of time. The CTP-HI_{low} represents here the atmospheric leg of the coupling and the SM is the contribution from the land. While the CTP-HI-SM joint probability space represents the average condition of the land-atmosphere scenario, the SM anomaly clearly shows that a patch of land which used to have dry coupling earlier might show a completely different set of characteristics (say, wet coupling) now; and thus; evolves temporally. This gives space for the incorporation of SM memory into this index as well as for drought recovery and persistence.

However, to increase the sensitivity of the test, the CDI devised by Roundy et al. (2006) has been modified. Instead of calculating over diurnal values individually, calculations have been done in two steps. First step includes considering a temporal period of 90 days other than one day done by Roundy et al (2006). Diurnal values of each parameter have been taken and average values have then been calculated over a period of 90 days (the March-April-May [MAM] period) to study yearly persistence and variability. Then, the rest of the calculation is done using a 3 month (MAM) mean.

Another main modification is that instead of applying the KS statistical test to the SM anomalies to determine whether they follow the mean distribution or not, the AD test is applied to make the results more accurate and precise. This is because the AD test gives more weight to the tails than the KS test, and we are more concerned with values which have deviated from the mean. Under drought conditions over India, the distribution of precipitation is less than two standard deviations below the mean. Due to land-atmosphere coupling and SM memory, it follows that the distribution of SM is also affected by precipitation and is deviated away from the mean. So, while testing SM anomalies over long temporal scales, the AD statistical test, which gives more weightage towards tails will give more precise results than the KS test, which focusses more on mean values.

Let us consider a test statistic undergoing the AD test. Since the underlying assumption in the test is that all the test statistics follow transitional coupling, and if any test statistic follows the mean curve, then it follows that the statistic has no preference for dry or wet coupling. Since, our main goal is to find areas in the CTP-HI_{low} space which are predominantly wet or dry, once the whole CTP-HI_{low} space is classified into dry and wet, we can create a time series of the number of dry coupling days (N_d) and, the number of wet coupling days (N_w) and build the Improved Coupling Drought Index:

$$CDI_{imp} = \left(\frac{N_d - N_w}{N_t}\right); \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots.. \text{(Equation 3)}$$

where, N_t is the total number of days in the time series constructed. Here, N_t , rather stands for the total number of events in the time series constructed, as MAM values are considered. The same applies for N_d and N_w.

A negative value near -1 indicates consistent wet coupling or drought recovery; whereas a positive value near +1 indicates dry coupling or drought intensification or persistence.

The prediction regarding drought persistence for the specified locations is finally cross validated using precipitation data and the standard precipitation index (SPI). We can then check if the performance in predicting drought persistence of CDI_{imp} is indeed improved and assess its accuracy.

Chapter 3: DATA SOURCES

For this study, one requires air temperature (T) and specific humidity (q) at atmospheric heights (z) equal to 1000hPa, 950 hPa and 850 hPa along with the values for surface pressure (z=1000hPa), air temperature at 2m above the surface (2m temperature), and specific humidity at 2m above the surface (q2m) to build the CTP-HI_{low} indices. Since hourly q2m values are difficult to obtain, an average of daily maximum and minimum q2m values are calculated and used. Along with these, daily SM estimates for volumetric soil water layer 1 (a depth 0-7cm) are also collected and averaged as it is required to complete the CTP-HI-SM matrix. All these data are for the period MAM for 1979-2010 at the location of this study.

This study focusses on west central India; which is one of the five most meteorologically homogeneous locations as characterized by the Indian Meteorological Department (IMD). It constitutes of the states of Telengana, Chhattisgarh, Maharashthra, and Madhya Pradesh of India. This is the region within the mainland of the Indian subcontinent defined by the box: (20°N, 74°E) to (25°N, 80°E) and (18°N, 84°E) to (13°N, 80°E). For ease of computation; this is loosely defined by the box: (13°N; 85°E; 25°N; 71°E) throughout this study; and for which all the data are calculated, manipulated and displayed. Figure A below shows the study area.

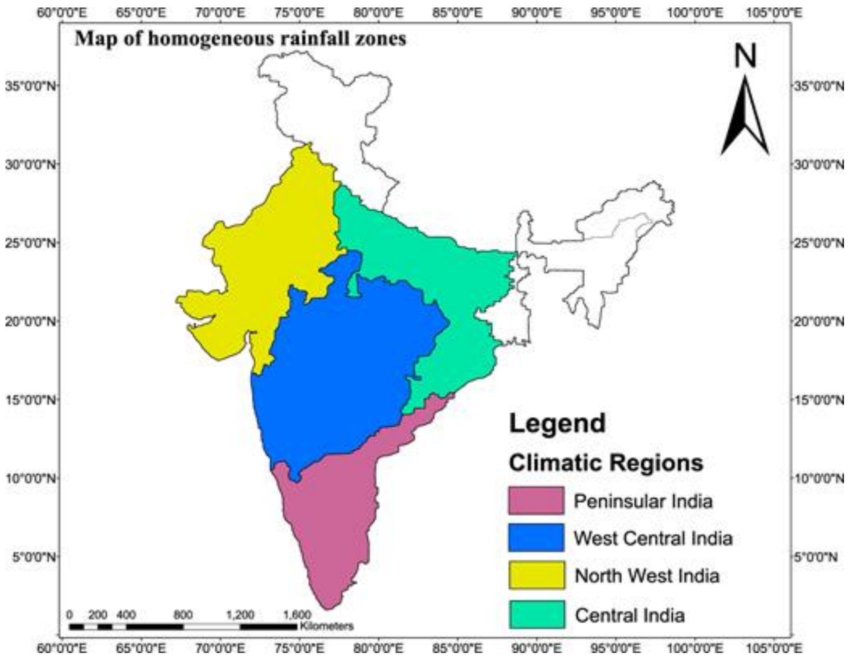

Figure A: Climatic regions of India, showing west central India as well (Syed et al, 2017)

Since transitional coupling conditions are suited for this kind of study as already discussed earlier, the time frame of our data is for the months of monsoon onset; MAM. The land-atmosphere parameters are studied from the years 1979 to 2010 owing to lack of data in periods prior to this. Total precipitation data are also collected for this period. These data are required to check the accuracy of the developed index regarding drought persistence or recovery.

The main source of data is from reanalysis models. A grid to grid coupling is considered to avoid spatial inconsistency. Initially, SM estimates were downloaded from Modern-Era Retrospective Analysis for Research and Applications version 2 (MERRA2); but due to data inadequacy over the study area; a new daily subset for the uppermost surface SM estimates (of depth 0-7cm, used for the final study) has been downloaded from European Centre for Medium-Range Weather Forecasts (ECMWF) climate reanalysis version 5(ERA5). Temperature and specific humidity estimates, both at surface and at 2m; along with surface pressure are downloaded from National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CSFR) for years 1979-2010 and from NCEP Climate Forecast System Reanalysis version 2 (CFSv2) for year 2011. All these estimates are mainly used for calculating the CTP. For temperature and pressure values at 2m, the 700 hPa atmospheric pressure level is considered. The remainder of the data are downloaded from ERA5.

The spatial resolution of the data for MERRA2 is 0.667°x0.5°. The choice of using MERRA-2 for data collection is not only its enhanced use of satellite observations, but also the way the project tries to include more aspects of the Earth system makes it a more reliable data collection resource. For example, one major development from MERRA to MERRA-2, is that, over land surfaces, MERRA-2 uses observation-based precipitation data as a forcing function to drive the land surface water budget (Reichle et al, 2016). This approach is like the gauge-based precipitation forcing developed for MERRA-Land. The precipitation forcing data derived from this approach is archived as the output variable called PRECTOTCORR in the MERRA-2 surface flux diagnostics (FLX) and land surface forcings (LFO) collections. We should also consider that the forcing precipitation is not purely based on only gauge observations, as it tapers back to MERRA-2 model generated precipitation poleward of 42.5° latitude and is completely MERRA-2 precipitation poleward of 62.5°. These are the places where there is a lack of observational data; so, the model generates precipitation data solely based on interpolation and climatology. Also, over continental Africa, the observations change to the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) gauge-satellite product. Care must be taken in mass balance studies as the difference between the observation-based and model-generated precipitation affect the water budget in land-atmosphere interaction studies.

ERA5 is the fifth generation of ECMWF atmospheric reanalyses of the global climate, and the first reanalysis produced as an operational service. ERA5 utilizes the best available observation data from satellites and in-situ stations, which are assimilated and processed using ECMWF's Integrated Forecast System (IFS) Cycle 41r2. The ERA5 HRES atmospheric data have a resolution of 31km, or 0.28125 degrees, and the ensemble of data assimilations (EDA) has a resolution of 63km or 0.5625 degrees. The wave data, however, are produced and archived on a different grid to that of the atmospheric model, namely a reduced latitude/longitude grid with a resolution of 0.36 degrees (HRES) and 1.0 degrees (EDA).

The temporal resolution of HRES data is hourly. The short forecasts have hourly steps from 0 to 18 hours. For the EDA, the sub-daily non-wave data (not used here) are available every 3 hours but the sub-daily wave data are available hourly.

The National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR) initially completed its 31-year period from 1979 to 2009 which was then extended till March 2011. The time series products at NCEP have been created at hourly resolution by combining forecasts and analysis products for each initialization time. NCEP upgraded their operational Climate Forecast System (CFS) to version 2 on March 30, 2011, after CSFR was discontinued. CFSR used to and CFS is initialized four times per day (0000, 0600, 1200, and 1800 UTC. CFSv2 monthly atmospheric, oceanic and land surface output products are available at various spatial and temporal resolutions; as low as 0.3-degree horizontal resolution and 6-hour temporal resolution. Here, however, daily estimates are used.

Chapter 4: METHODOLOGY

At first, the time period is set as the months of March, April and May (MAM) for each year (1979-2010) for each of the parameters over the region of study loosely defined by (13°N; 84°E; 25°N, 71°E).

Then, the mean value of soil moisture (SM) over the entire temporal period and spatial region is calculated from diurnal values. Using this mean value, anomalous values for each time period (MAM of each year) are calculated. Thus, unlike the Coupling Drought Index (CDI) developed by Roundy et al (2012), where the matrices formed used percentile values of soil moisture at daily time steps, I use the soil moisture anomaly values at a temporal period of three months, although the calculation is started by taking diurnal values. Thus, no information is lost while the calculation is done, but information is compressed to reduce the number of matrices and the ease of numerical manipulation. This is because the period under study is large as compared to the study conducted by Roundy et al (2012).

Note that, the outliers are detected using the mean values rather than the median values. This is a significant modification done to the original index developed by Roundy et al (2012). This is because, when a large time period is considered, although there are deviations from the mean, the data set is not that skewed over the study region impacted by droughts. When I infuse daily, seasonal and climatological means into the calculations to get the anomalous values; I consider not only statistics, but also the physical and geographical factors affecting SM in the region under study.

These values thus obtained; are unique and are the contribution of the land in the land-atmosphere coupling regime. These also contribute to the soil moisture memory, which forms an integral part in the prediction of drought persistence or recovery. This is another reason why SM anomaly values are calculated here over a period of three months (MAM). Although daily values of SM anomalies help us detect triggering of convection and thereby precipitation; seasonal values of SM anomalies, when calculated using diurnal values, help us in putting the contribution of memory in addition to the land-atmosphere feedback mechanisms that drive precipitation at diurnal time scales.

Thus, for building a predictive coupling index which predicts long-term seasonal occurrences like droughts; the time period for building the joint probability space using the predictive components is also chosen and built like this. For calculating the other components, viz. Convective Triggering Potential (CTP) and Humidity Index (HI) as well, we start off using daily values to finally build a matrix with a temporal period of three months; where each time step (or a row) is designated by a year from 1979 to 2010; such that the distribution is continuous. It is important for the distribution to be continuous for the data to qualify for the application of the

Anderson-Darling Test (AD test). This is significant; as the SM anomaly dataset is skewed; and hence, AD test is applied instead of the Kolmogorov-Smirnov statistic test (KS test) as done by Roundy et al (2012); since, AD test gives more weight to the tails than compared to the KS test. It is another significant modification; as in the index developed by Roundy et al (2012); the skewness is removed in the beginning of classification or building of matrix; while, here; the process of matrix formation or classification of the land-atmosphere parameters are dependent on the skewness of the data.

The SM anomaly values are calculated as percentages or fractions; i.e.; the deviations are calculated as fractions of the mean soil moisture estimate as calculated over the entire time period. Then the percentiles of these fractional values are taken. So, the values range from 0 to 1. This is how they can be incorporated easily into the joint CTP-HI-SM matrix; based on their numerical values. It also becomes easy to plot the distribution of the SM anomalies as soon as they as they are represented in this way.

For building the CTP-HI matrix; we need to individually calculate CTP and HI.

For calculating the HI, at first, I downloaded the values for specific humidity (q) at the atmospheric pressure levels (p) of 850hPa and 950hPa. From this, I calculate pressure levels (e) of water vapor component of the air using the psychometric constant:

$q = \gamma\, e/p$ …………………………… (Equation 4a) or, e = pq/ɤ, where ɤ = 0.622 is the psychometric constant ……….. (Equation 4b)

Using the values of vapor pressure (e), we calculate dew point temperature at those pressure levels using the following formula: (Equation 5)

$$T_{d=}\frac{\ln(e)+0.49299}{0.0707-0.00421\ \ln\,(e)}$$

Thus, I have dew point temperatures for 850hPa and 950hPa atmospheric levels. The unit for the dew point temperature is in Kelvin (K).

Next, I download the values for atmospheric temperatures at these two atmospheric pressure levels. The downloaded data (in K) has two grids: the generic grid and the latlon grid. The latlon grid has the values of the data in terms of latitude and longitude; with the pressure as the z-coordinate; in terms of Cartesian coordinates. I choose the latlon grid for our convenience and then regrid the data to 0.50°x0.50°grid size. In this way, the grid sizes of the data for dew point temperature and the atmospheric temperature match. This is going to help the study to compute the Humidity Index (HI_{low}); when I add and subtract values within the same grid. For calculating HI_{low}, I use the following formula: (Equation 2)

$$HI_{low}=(T_{950}+T_{d950})+(T_{850}+T_{d850})$$

HI_{low} is in Kelvin(K). Once it is computed, the selected time period and the selected spatial region is selected. Then, HI_{low} is also converted to °C.

Next, I calculate the Convective Triggering Potential (CTP). One must remember that, CTP is the integral of the curve between the moist adiabat and environmental lapse rate from 100mb above the ground to 300mb above the ground. So, CTP is derived as the integrated space between the environmental temperature profile and a moist adiabat from 900 mb to 700 mb. Performing linear interpolation makes it possible to extract each value at the desired levels between 900 mb and 700mb following the temperature profiles. This is done by finding the y-intercept equal to the pressure level minus the desired level; i.e.; one basically finds the temperature and dew point corresponding to the level.

At first, I calculate the Moist Adiabatic Lapse Rate (MALR) using the following formula: (Equation 6)

$$MALR\ (\gamma_m)\ =\ \gamma\ -\ \frac{\lambda}{c_p}\frac{\partial q_{sat}}{\partial z};$$

where, environmental lapse rate (ELR) = $\gamma = -\,\partial T/\partial z = g/c_p$;

with acceleration due to gravity = g = 9.81 m/s^2; and

specific heat of air at constant pressure = c_p =1.005kJ/kgK;

latent heat of vaporization = λ = 2.5e6 kJ/kg;

difference between the saturated specific humidity at two specified levels = $\delta q_{sat} = q_{sat}$ (900mb) - q_{sat} (700mb);

difference between the pressure levels marked as the z-coordinate = δz_t = 900mb-700mb = 200mb

Putting the values for saturation specific humidity at 900mb and 700mb; one gets the values of the Environmental Lapse Rate and the Moist Adiabatic Lapse Rate.

Convective Triggering Potential (CTP) is calculated using a simple formula; (Equation 1b)

$$CTP = CTP + R_d \left(t_{par\_mid} - t_{seg\_mid}\right) * log({}^{p_{old}}/{}_{p_{seg}})$$

where, par stands for parcel,

seg stands for segment,

mid stands for middle.

Broadly, this talks about the rising air parcel following the lapse rate. The first measurement is taken 100mb above the ground level, i.e., at 900mb pressure level; and the last measurement is taken 300mb above ground level; i.e., at 700mb pressure level. Since, one takes several measurements following the air parcel at different pressure levels throughout the planetary boundary layer in between these two levels; each of these smaller hypothetical homogeneous divisions within the PBL are called as segments. For ease of calculations, often, the temperature and pressure are noted at the middle of these segments.

As each value at a comparatively higher level following the air parcel is considered; it is calculated using the values at its immediate lower level. The values at the lower levels are then referenced as “old”; and comparatively newer values are updated at each step of the calculation. This applies for temperature, pressure, humidity profiles as well as the overall CTP calculation. In the equation for CTP, however, only p_{old} is mentioned explicitly.

So, as already discussed; here, the temperature (t) profile is taken from the Environmental lapse rate. Since, the pressure can be indicated by the MALR, this equation can also be simply understood with the help of the skew T-log P diagram; where same relationships are used to study atmospheric thermodynamics.

Namias (1940) had proclaimed that the “best” isentropic surface to diagnose low-level moisture and vertical motion varies with latitude, season and the synoptic situation. For summer, the preferable surface is from 310-315K while from Spring it is from 295-300K. Here, for the MAM months i.e., the pre-monsoon period in India; which predominantly marks the spring to summer period; manifested with an abundance of moisture in the air; looking into the skew T-log P diagrams over the study area; one can choose a surface 300-320K for further isentropic analysis.

The initial CTP on the right-hand side is taken as zero, for we do not want to consider any other CTP calculations below 900mb.

R_d stands for the molar gas constant for dry air. Dry air is taken into consideration because, initially, the air parcel rises with the dry adiabatic lapse rate (DALR); till, finally the profile merges into the environmental lapse rate. We know that, in general, the value of the ELR lies between the DALR and the MALR; i.e.; DALR>ELR>MALR. For the chosen levels, i.e.; at 700mb and 900mb (or at 300K and at 320K); ELR>MALR.

The next steps include calculating temperature, pressure and humidity profiles for each of the segments; i.e.; t_{seg}, p_{seg}, q_{seg} and thereby calculating t_{mid}, p_{mid}, q_{mid} and updating t_{seg_old}, p_{seg_old}, q_{seg_old}.

For the segment calculations, initially, for t_{seg_old}, p_{seg_old}, q_{seg_old}, I consider their values at 900mb; and then keep updating those values for consecutive higher layers. After finding pressure increment between defined levels; and thereby finding p_{seg}; for t_{seg}, q_{seg}; generally, linear interpolation is done to get temperature and specific humidity profiles at the increments. However, here, to get better accuracy, I have chosen six equally spaced pressure levels, viz., 750mb, 775mb, 800mb, 825mb, 850mb, 875mb; between 700mb and 900mb to get seven segments; for which temperature and humidity readings are taken (and not interpolated) at each pressure level.

For tmid, pmid and qmid; I use the following set of formula: (Equations 7a, 7b, 7c)

$$t_{mid} = \frac{(t_{seg}\log p_{seg} + t_{seg\_old}\log p_{old})}{\log(p_{seg} * p_{old})}$$

$$p_{mid} = \frac{(p_{seg}\log p_{seg} + p_{seg\_old}\log p_{old})}{\log(p_{seg} * p_{old})}$$

$$q_{mid} = \frac{(q_{seg}\log p_{seg} + q_{seg\_old}\log p_{old})}{\log(p_{seg} * p_{old})}$$

where, the value of p_{old} for each layer at each iteration is given by p_{seg}; i.e.; the pressure level of the previous segment already calculated. So, again, since the calculations start from the 900mb pressure level; and the values for that level are initialized for carrying out the iterations; the initial value of p_{old} is also 900mb. Here, in this case; one needs to keep in mind that the first set of t_{mid}, p_{mid}, q_{mid} are calculated for the 875-900mb pressure level; where, p_{seg} is considered as 875mb and p_{old} as 900mb. For the next levels; of course, the values are updated accordingly.

Next steps are calculating the temperature profiles for the air parcel given by the following variables: t_{par}, t_{par_old} and t_{par_mid}.
Again, for the initial t_{par_old} is taken as t(900mb) i.e., the temperature of the air parcel at 900mb as given by the ELR. We keep updating t_{par_old} as the air parcel keeps rising.

Now, t_{par} is given by, (Equation 8)

$$t_{par} = t_{par\_old} - MALR * dz$$

where, pressure difference = $dz = (p_{old} - p_{seg})/(grav * p_{mid}/(R_d * t_{mid} * \left(\frac{1+{q_{mid}}/{e_p}}{1+q_{mid}}\right))$

with, acceleration due to gravity = grav = 9.81m/s^2;

psychometric constant = e_p = 0.622; and

molar constant of dry air = R_d = 287.04 kg/K

The resulting value of dz is then divided by 1000 to maintain consistency with units and keep the final resulting value in mb.

Then, t_{par_mid} is given by, $t_{par\_mid} = 0.5(t_{par} + t_{par\_old})$ ………… (Equation 9a)

and similarly; , t_{seg_mid} is given by, $t_{seg\_mid} = 0.5(t_{seg} + t_{seg\_old})$ ….. (Equation 9b)

So, thus, Convective Triggering Potential (CTP) is calculated for each pressure level (or segment) given by the formula (already written earlier): (Equation 1b)

$$CTP = CTP + R_d \left(t_{par\_mid} - t_{seg\_mid}\right) \log({}^{p_{old}}/_{p_{seg}})$$

It is important to note that; this formula helps in calculating the CTP for each layer individually and then adding the CTP for the higher-pressure level segment with the already calculated CTP for the lower layers. Finally, this is how I get the CTP for the entire pressure layer: 900-700mb.

This gives the values for CTP, HI & SM anomalies for the period of MAM from 1979-2010 over South west India. With these three sets of values; one can easily form a joint CTP-HI-SM space. So, now; I consider the two-dimensional CTP-HI space having (n x n) bins and I try to classify each bin based on the marginal distribution of soil moisture.

The distribution of the marginal variables (the marginal distribution) is obtained by marginalizing; i.e.; focussing on the sums in the margin; over the distribution of the variables being discarded; and the discarded variables, then, are said to be marginalized out. To build the 2-D CTP-HI space of (n x n) bins from the 3-D CTP-HI-SM space; I apply this same mechanism; such that, each bin consists of a marginal distribution of SM; i.e.; the values of SM are incorporated in such a way, that; in the final joint probability space of CTP-HI; SM anomalies are marginalized out.

Next, with the help of the newly formed (n x n) 2-D CTP-HI matrix; I find out the Cumulative Distributive Function (CDF) for each bin. By definition, CDF of a function X of x is the probability that the variable takes of values less than or equal to x;

$$\text{i.e.; } F(x) = \Pr(X \leq x) = \alpha$$

So, at this point of the calculation; one is left with only two sets of values; the bin SM anomalies and the values of CDF for the bin marginalized 2-D CTP-HI space for each bin. It is known that, the total number of bins is taken as n. This is equal to the number of years under study. With these sets of values, we perform the Anderson-Darling (AD) test on our data for normality.

If H_0 is defined as the null hypothesis; where, the data to be tested follows a specific distribution (mean distribution in this case) and H_a is defined as the alternate hypothesis; where, the data do not follow the specified distribution.

The test statistic is given by, $A^2 = -N - S$; where,

$$S = \sum_{i=1}^{N} \frac{(2i-1)}{N} \lfloor \ln F(Y_i) + \ln\{1 - F(Y_{N+1-i})\} \rfloor$$

where, F is the Cumulative Distributive Function (CDF) of the specified distribution,

Y_i is the ordered data,

N is the total number of data tested. ……….. (Equation 10a)

So, it is a very sensitive test, more sensitive than the Kolmogorov-Smirnov test; giving more weight to the tails. But its disadvantage lies in the fact that the critical values must be calculated for each distribution. The test is a one-sided test and the hypothesis that the distribution is of a specific form is rejected if the test statistic, A; is greater than the critical value. For a given distribution, the A-D statistic may be multiplied by a constant (which usually depends on the statistic, n). The needed constant is typically given with the critical values; which in turn are dependent on the specific distribution that is being tested. These critical values also determine the critical region at a particular significance level α at which the data is tested.

The CDF is the associated cumulative distributive function and invCDF is the associated inverse of this function for any Probability Density Function (PDF). Now, let, Y (=y_1, y_2,.., y_n) be a sample which is ordered in X (=x_1, x_2, .., x_n). The series P (=p_1, p_2,.., p_n) defined by p_i = invCDF(x_i) and Q (=q_1, q_2,.., q_n) defined by q_i =invCDF(y_i); where, P is the unsorted and Q is the sorted array. The samples are drawn from a uniform distribution only if Y (and thus, X) are samples with PDF. At this point, the order statistics are used to test the uniformity of P (or for Q), and for this reason; the values of X are ordered (in Y). On the ordered probabilities (on P); several statistics can be computed; and AD is one of them.

$$AD = AD\ (P,n) = -n - \sum_{i=1}^{n} \frac{(2i-1)\ln(P_i)\left[1-P_{(n-i+1)}\right]}{n} \quad \text{(Equation 10b)}$$

The associated AD statistic for a 'perfect' union distribution can be computed after splitting the [0,1] interval into n equidistant intervals; i/n; with $0 \leq i \leq n$; being the boundaries; and using the middles of those intervals $r_i=(2i-1)/2n$

$$AD_{min}(n) = AD(R, n) = -n + 4\, H_1(R, n)$$

where, H_1 is the Shannon entropy for R in nats (unit of information or entropy). (Equation 10c)

The equation gives the smallest possible value for AD. The value of the AD increases with the increase of the departure between the perfect uniform distribution and observed distribution (P).

The critical values for upper tail level percentage α, n, modified A^* and A and their inter-relationships are taken from Stephens (1979).

For tests for normal distribution, for all all $n \geq 5$; the modified A^* is given as; (Equation 11)

$$A^* = A^2(1.0 + {}^{0.75}/_{n} + {}^{2.25}/_{n^2})$$

At 95% confidence level, α=0.05, p=0.752=75.2%

At 90% confidence level, α=0.10, p=0.631=63.1%

At 85% confidence level, α=0.15, p=0.561=56.1%

The null hypotheses that the random variable X has the distribution F(x;0) is rejected at level α if A exceeds the appropriate percentage point at this level. Here, the rationale is based on the idea that coupling is a recurring process and absence of data suggests an inherent randomness. Hence, I test for normality for n=10 to 32 and for α=85 to 95%.

For better accuracy, the test is repeated, replacing X_i with the maximum likelihood estimator (MLE), Z_i, of X_i. (Equation 12)

$$Z_i = F\big(x_i; \hat{\emptyset}\big) = C \prod (F(X_i) - F(X_{i-1}))\,(1 - F(X))^n$$

All bins accepted under the AD test are then classified as dry or wet. One must understand that this classification is dependent on the mean SM of the classified bins. If the soil moisture of a particular bin is more than the average soil moisture; then the bin is classified to be wet; whereas, if it is less than the average soil moisture, it is classified to be dry. For bins which cannot be classified as dry or wet, it is considered that, the coupling is transitional. Then, I try to conduct the AD test at a higher significance level to fit in more bins.

Thus, for each significance level; I shall have number of dry days (N_d), number of wet days (N_w) and total number of days (N_t). The total number of days varies for calculations for each significance level because, the number of bins increases as we increase the confidence level. So, for each significance level; we calculate the Improved Coupling Drought Index (CDI_{imp}) given by; (Equation 3)

$$CDI_{imp} = \frac{N_d - N_w}{N_t}$$

The CDI_{imp}, like CDI, has a minimum value of -1 and a maximum value of +1. More the value is near -1; it indicates consistent wet coupling, which, in turn, indicates, drought recovery. Similarly, a value near +1 indicates regular dry coupling or drought intensification. A value of zero, i.e.; when the number of dry coupling days equals the number of wet coupling days; it is assumed that there is no net effect towards or away from the drought. One should remember that, this index only captures the impact of precipitation and evaporation on drought. So, atmospherically controlled events can also drive the value towards zero; irrespective their actual contribution towards coupling and thereby drought intensification or recovery.

Chapter 5: RESULTS AND ANALYSIS

The data is tested at different significance levels to include more bins in the testing and make the results more accurate. We consider a sample size of 20 or less as insufficient to populate a distribution. In such cases we classify the bin to be atmospherically controlled. As coupling is a recurring process and the absence of data suggests an inherent randomness, so, we continue classification and testing at higher significance levels. At three different confidence levels, 24, 27 and 32 bins were considered, each bin signifying each year under study.

When we tested the data at 85% confidence level; the value of CDI_{imp} came as -0.0833333 or -0.08. This value is slightly negative; and signifies wet coupling or a miniscule chance of drought recovery. At 90% confidence level, the value of CDI_{imp} came as 0.037037037 or 0.04. This value is slightly positive, signifying dry coupling and slight chances of drought intensification. At 95% confidence level, the value of CDI_{imp} came as 0.00. This indicates that there is not much significant difference between the two distributions; and hence the transition of the atmosphere is closer to the climatology. So, this implies that there will be no net effect towards or away from the drought, as it is neither wet nor dry.

The next step is to check whether this result given by the index is accurate or not. To check this, I consider precipitation, more specifically daily mean precipitation over the study area as a parameter to identify drought recovery or intensification. I consider precipitation for the years 1979-2010, on which my study is based. For checking the accuracy of the index developed, I consider the precipitation for the next year; 2011.

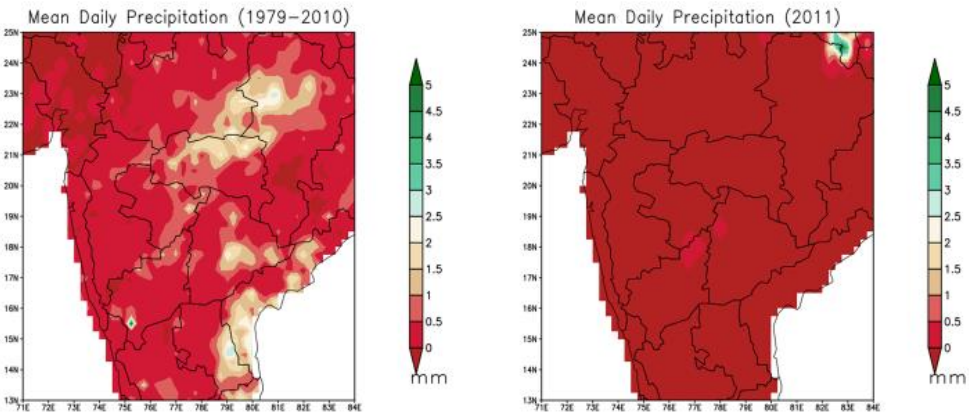

Figure 1: Mean Daily Precipitation (in mm) for the years 1979-2010 (left) and for the year 2011 (right) over South West India

For the drought period under study 1979-2010; the mean daily precipitation for that time frame is considered. Generally, majority of the area under study gets rainfall 0.5-1mm daily rainfall; i.e.; 18.25-36.5cm average yearly rainfall. This makes the area as one of the most severely drought affected areas; as the average precipitation in India is about 120cm; with the majority coming from the monsoons during the JJAS period;

and certain areas getting average yearly rainfall as high as 185cm. After hitting the Indian peninsula, the Indian summer monsoonal current generally gets divided in two branches; the Bay of Bengal (BoB) branch and the Arabian Sea branch; and each of its branches is responsible for precipitation at each of the zones in the country. So, it is but important to investigate the source of rainfall for the area under study. As the area lies in the south western part of the country; its proximity to the Arabian Sea makes the area under study a location which is under the influence of the Arabian Sea branch. That is why there are isolated separated locations where the mean daily precipitation is as high as 2.5mm; making the mean annual precipitation above 90cm; but such zones are sparse. This is because most of the study area falls in the rain shadow area of the Western Ghats. Other than this geographical factor, there are also other dynamical factors (like ENSO) & local factors (like irrigation & vegetation changes) that affect these rainfall patterns and hence, the drought conditions.

Looking at the mean daily precipitation for the years 1979-2010 (Figure 1), one can see that although most of the study area receives daily mean precipitation less than 1mm; some coastal areas at the southwest corner receives precipitation as high as 2.5mm/day. There are also isolated patches of 1mm/day at various locations, towards the north and at the mid-west. Looking at the year 2011 (Figure 1); one finds that again almost the entire study region has received a daily mean precipitation of 0-0.5mm; with two isolated patches near the center receiving marginally higher precipitation of 0.5-1mm/day rainfall. One isolated patch of heavy precipitation of 5mm/day is also noticed towards the north east corner of the study area. Overall; if precipitation records are to be followed; drought neither intensified nor was there any recovery; as predicted by the CDI_{imp}. So, the index is pretty accurate.

Somewhat similar pattern of intensity of precipitation is seen in the plots for Standardized Precipitation Index (SPI) for 1979-2010 (left) and 2011 (right) in figure 2 below. The SPI is a relatively new drought index based only on precipitation. It is an index based on the probability of precipitation for any time scale. SPI is generally used to identify meteorological droughts for a variety of timescales. For a shorter temporal scale, SPI is often associated with changes in soil moisture, while for longer temporal scales, SPI can help identify groundwater storage. In either case, it helps in quantifying the water content in the soil. It also quantifies observed precipitation as a standardized departure (say, standard deviation or variance) from a selected probability distribution function that models the raw precipitation data. The SPI generally uses monthly data as inputs and the output can be created for 1 to 36 months. Here, I have taken (downloaded) annual SPI values; i.e.; values for 12 months.

A 12-month SPI is a comparison of the precipitation for 12 consecutive months with the same 12 consecutive months during all the previous years of available data. The SPI at these time scales reflect long-term precipitation patterns. The value of SPI tends towards zero if no major hydrometeorological event is taking place.

However, SPI does not account for evapotranspiration, runoff etc and can give erroneous results if is calculated for less than 30 years span; as it is primarily developed for detecting abnormal wetness at different time scales or droughts; which does not show up in lesser time scales than that. Hence, for the study; I just consider the pattern showed by the SPI for 2011; but the detailed intensity of precipitation is ignored.

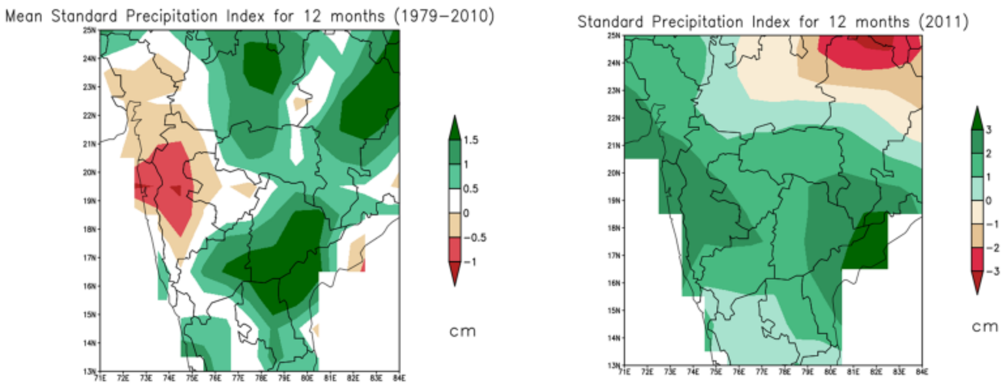

Figure 2: Mean SPI for 12 months for 1979-2010 (left) and for 2011 (right)

It is to be noted that, the same areas remain dry or wet in terms of precipitation as figure 1. So, though, the study incorporates SPI; the main indicator for drought identification, persistence and recovery; to validate the study; especially for smaller time scales remains precipitation. The figure 2 also shows that at least for smaller time scales, one needs a more accurate predictive index for drought.

The improved coupling drought index developed uses the main underlying concept of land-atmosphere coupling; so, I try to understand the various components of the index and their respective importance in the development of the index; i.e.; how they help in the prediction of the drought over the study area. If I concentrate only on the data in the MAM time period from 1979-2010 based on which the index is developed; then I can try to understand the role of land and atmosphere roughly on the index; i.e.; by studying the mean Humidity Index and the mean Soil Moisture data for volumetric layer 1; and comparing them with the mean total precipitation; I try to develop an understanding of the atmospheric leg and the lithospheric leg and their influences on the development of the index.

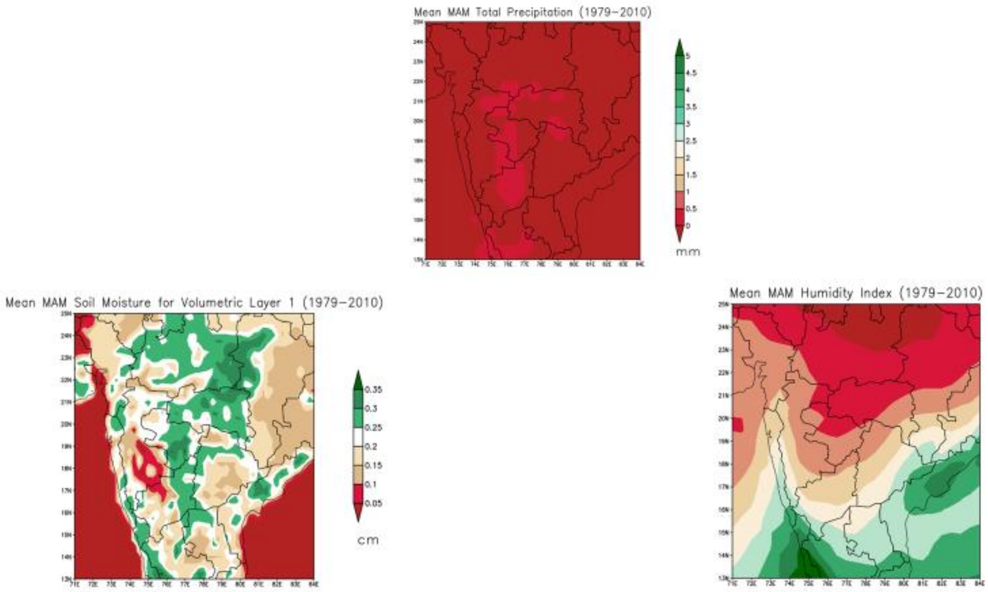

Figure 3: Factors affecting and influencing the coupling index: (from top, anticlockwise) Daily Mean values for Precipitation, Soil Moisture for volumetric Soil Level Layer 1 and Humidity Index for MAM period for 1979-2010 (study period) over South West India (study area)

As seen in Figure 3; for the MAM period from 1979-2010; most of South West India receives no precipitation (0 cm). Some parts, the south west corner of the study area, which lies near the sea, and the central region receives marginally higher rainfall of 0.0001cm/day. The wettest region lies at the middle of these two areas in the center, one towards the middle east of the study area and the other towards the central south; where precipitation is as high as 0.0006cm/day. The mean MAM humidity index plot, on the other hand, looks quite uniform; with places away from the sea towards the north having greater values of HI; with northernmost landlocked areas at the center of the study area having values as high as 350K; and coastal areas having lower values; ranging from 0-150K. The mean MAM soil moisture plot for volumetric layer 1 is very scattered; with values ranging from 0.05 to 0.35 cm. The wettest region falls in a location where the soil is the driest. The soil is the wettest in some coastal areas; where; humidity index is as low as 0K and some inland areas where there is no precipitation and humidity index is as high as 350K.

This can be explained by the fact that the humidity index identifies areas of saturation deficit in the plots. The behavior of HI is linked to the classification of the CTP-HI space; i.e.; by the general separation of the wet and dry spaces in the HI climatology. Findell and Eltahir’s classification was based on days characterized by triggering of afternoon convective precipitation while Roundy had considered all days. The study region, however, is under the influence of the monsoons. So, this study is dealing with a particular subset of Roundy’s classification; where there are some modifications or extra atmospheric conditions persisting.

Reduced sample size weakens the spatial extents; but the robustness in general patterns are still quite visible. One can visually see this by looking at the daily mean precipitation plots for 1979-2010 for the whole year versus only for the months MAM based on which our index is developed (Figure 4, below).

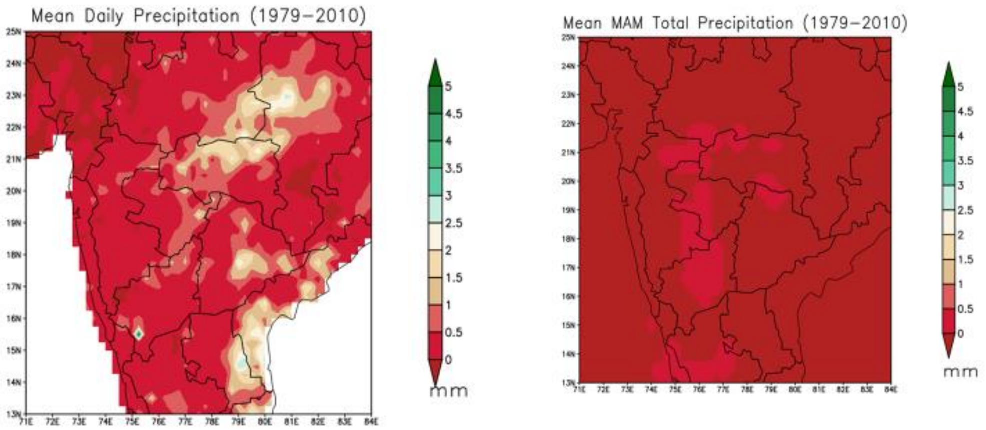

Figure 4: Comparing mean daily precipitation over South west India for the whole year (left) and the MAM (study period) for 1979-2010 over South west India

Since the wettest area in terms of precipitation is also the driest area in terms of soil moisture; and some of the most driest areas in terms of precipitation are also really wet in terms of soil moisture; it is clearly evident that both dry coupling and wet coupling happen over the study region. At different locations, due to the pre-monsoonal conditions prevailing; they also act as extra external factors to make the coupled system reach equilibrium. For the system to reach wet equilibrium, precipitation occurs; and for the system to reach a dry equilibrium; evaporation occurs. The determination of such a coupling event; and thereby the occurrence of precipitation and evaporation is mostly controlled by the HI. If one looks closely, majority of southwestern India has a very high value of humidity index of above 150K/day. These areas are the same places which receive almost zero precipitation and are extremely dry. The overlapping zones of low humidity index and high soil moisture and the reverse is already stated above.

Next, I try to break down the dry and wet coupling events and their relationships with the main variables. When it comes to coupling of soil moisture, and the development of the index; then, if one takes a closer look at the building of the index with dry coupling and wet coupling days, then, as it can be seen from table 1, the number of events for dry coupling and wet coupling are equal.

Type of coupling event	Number of events	Sum of CTP (J/kg) by coupling	Sum of HI (°C) by coupling
Dry	16	4457.298	-829.786
Wet	16	4490.980	-748.614

Table 1: Table showing the number of dry and wet coupling events, the dry and wet coupling events in terms of CTP (J/kg) and in terms of HI (°C)

But when it comes to analyze dry and wet coupling in terms of CTP (Table 1) and HI (Table 1); the quantitative values give a clearer picture. For both CTP and HI; the quantitative value for the wet coupling events are higher; i.e.; both the values of CTP and HI are more on wet coupling events than compared to dry coupling events.

If one can understand these coupling mechanisms and these feedbacks; then that would help us in understanding the past climate as well as the future projections. A positive feedback or wet coupling signifies recovery. In case of positive feedback, there are more chances of occurrence of precipitation on a wet land surface and evaporation is not limited just by the available soil moisture. The latent heat is released into the atmosphere; the moisture flux thus released increases the specific humidity of the atmospheric boundary layer (ABL). When this moisture rises to layers of conditional instability; convectional precipitation may occur; further increasing the soil wetness; and thus, enhancing the positive feedback.

However, for dry land surfaces; the land surface limits evaporation by moisture availability. The smaller moisture flux is insufficient to induce convection; and hence, no precipitation occurs; keeping the land surface dry as before; thus, making the drought conditions persist, or, in cases of longer time periods, to intensify. This is another case of positive feedback; but, in this case, the drought intensifies.

However, in case of negative feedbacks, precipitation occurs over dry land surfaces and wet land surfaces largely remain dry. For the area under study, one can observe that; negative feedback is occurring over most of the extent of South Western India. Generally, this happens when some potential instability develops above the ABL top. Over dry land surfaces, a larger value of sensible heat flux is observed. This often gives rise to the growth of a larger ABL. Whether this ABL would trigger convection or not depends on whether this ABL is able to entrain the stable layer and reach the unstable layer above or not. Since, the CTP and HI gives an idea of the atmospheric profile of the rising air parcel; therefore, knowledge of the CTP-HI framework is crucial to the study. If the surface conditions are wet for a dry soil; i.e.; there is a saturation deficit; which can be indicated by a high value of humidity index; then there would be no convection; and hence, no precipitation. The same is observed over the study region; especially over the regions where the impacts of negative feedback are the strongest.

Overall, however, if one looks at the persistence of each coupling event, i.e.; how long, each coupling event, i.e.; persistence of MAM mean from one year to other, dry or wet, lasts; it is found that the wet coupling events persists for a little bit longer (by 15% more) time than compared to dry coupling events (Figure 5).

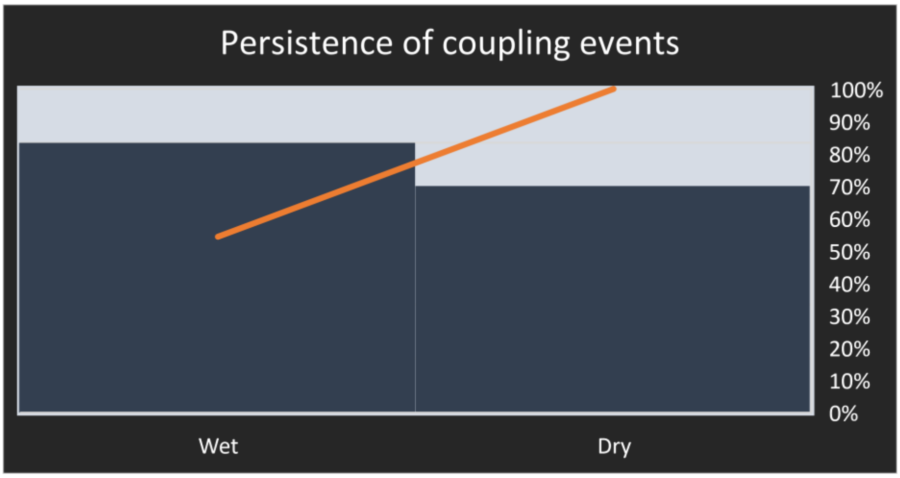

Figure 5: Persistence of Coupling events (in terms of dry and wet events)

A year to year analysis can be found in figure 6; which shows that the pattern is quite erratic. But the only thing common is that the system always wants to come to an equilibrium. It is to be remembered that, while building this index; such events are marked out using soil moisture anomalies. So, the distribution of SM anomalies during MAM (1979-2010) (Figure 7) must mirror Figure 6.

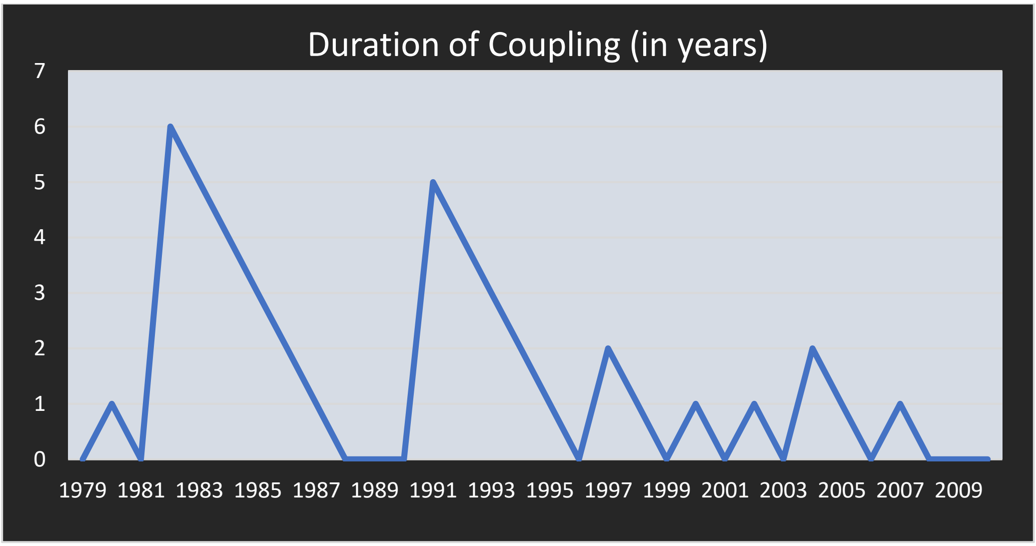

Figure 6: Duration of each coupling event (in years)

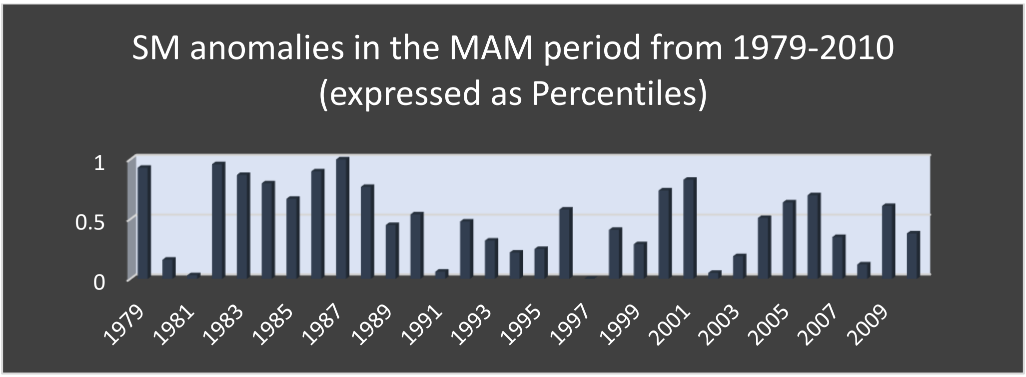

Figure 7: Distribution of Soil Moisture Anomalies for MAM period (1979-2010) expressed as percentiles

Hence, soil moisture plays a very important role in the persistence of the event; i.e.; soil moisture controls the memory of the coupling and the drought. One can understand this because there is not much variability in the atmospheric variables during a dry or wet coupling event, but distinct differences in precipitation and soil moisture. So, increase or decrease in the amount of initial soil moisture can control the duration of coupling; and thereby, the duration of the drought event. However, this change in SM cannot act as a trigger for the coupling event in general.

Therefore, humidity index controls the spatial aspect, i.e.; where there is a saturation deficit and hence a coupling event might occur; and the soil moisture controls the temporal aspect of the coupling event; i.e.; for how long the coupling event would last; depending on the initial SM. Thus, they control different aspects of the drought. Since, this is how the components of the index influence the index, and more importantly a coupling event; the index, with slight modifications, should be useful in studying other hydrometeorological events where land-atmosphere coupling is involved as well.

Chapter 6: DISCUSSIONS AND CONCLUSIONS

The improved coupling drought index (CDI_{imp}), which is developed in this study, is an alternative predictor of droughts to the already existing drought indices and it is applied to west central India; a tropical zone, where, monsoon dynamics play a key role in determining the precipitation. It takes into consideration the land-atmosphere coupling and the role of the ABL to detect convection and moisture in the atmosphere to predict if there is going to be rainfall or not. This, in turn; helps to know whether drought will intensify, continue or recover.

However, the study accurately predicted that the drought conditions will persist (neither intensify nor recover) for 2011 after studying pre-monsoonal conditions for the period 1979-2010. Using precipitation data to cross validate, the prediction was found to be accurate over the study area.

The main factors in developing the index are primarily the Convective Triggering Potential (CTP), Humidity Index (HI), and the Soil Moisture (SM) for the upper most layer of the soil. The former two represents parameters of the atmosphere, more specifically the atmospheric boundary layer (ABL). It was found that, HI, which helps to find the saturation deficit of an air parcel; is crucial to identify the spatial aspect of a region affected with drought. The SM, however, plays the role of identifying the temporal aspect; since it is associated with the memory or persistence of a hydrological event. So, the HI determines where a drought would or would not occur; whereas the SM determines for how long it would occur. The CTP, as the name suggests; acts as a trigger for a hydrometeorological event to begin. The manifestation is seen in precipitation.

The study, although gives quite accurate results; suffers from some inherent flaws. The study could not be carried out till 2020 due to lack of data of certain parameters over the Indian subcontinent. The precipitation and the SPI data used to validate the study results are in fact produced by combining gridded observed and reanalysis precipitation estimates. This is done because no existing dataset is ideal. Observed datasets are the closest to "ground truth"; but are uneven in quality and coverage. Radiosonde observations are not available throughout the study area; and following the atmospheric profile. The most accurate products are available on time intervals and are sometimes too long for the study purposes (monthly, for example). Model precipitation, for example, has complete coverage and is available at sub diurnal time steps; i.e., it mirrors time-series of original precipitation well; but; atmospheric state variables are constrained by data assimilation, typically strong time-mean and diurnal biases. Additionally, while combining data from 3 different model reanalyses; although regridding has been done to minimize errors that arise due to different resolutions, still the results are not free from different systematic biases. Future studies should definitely probe into this aspect.

Generally, dry coupling events are associated with continuation and intensification of drought (drying of deep soil layers) while wet coupling is associated with wetting of deeper soil layers; and hence drought recovery. Thus, if one observes multiple events within a time frame, that can be indicative of hydrological extremes of any kind. So, this index can be used to detect both drought and flood.

For India, during January to May; the predominant flow is from the north; bringing dry and cool conditions. By the end of May; the land surface gets heated by increased solar radiation; and this causes rising air masses over land. This heat low over the land draws the moist oceanic air towards it; bringing about the summer monsoon onset in June. Generally, the monsoon rainfall continues for the period JJAS, and constitutes most of the annual precipitation over the Indian subcontinent.

During the period and along the region of study, however; effects of orography, wind shear, and synoptic systems that affect atmospheric conditions are also relevant for the land-atmosphere interaction; but are omitted. A perfectly mixed ABL, constant soil moisture and cloud free conditions are also assumed. The importance of local feedbacks from land surface to convective precipitation has been previously quantified for India using the CTP-HI framework (FE2003a); but; has not been discussed here.

If yearly CTP-HI cycle can be studied; then one might also observe the signature of the monsoon dynamics over the study area; as studied by Tuinenberg et al (2010) over India. However, earlier, Tuinenberg et al. (2010) had stated that although the CTP-HI framework works well in determining regions which are potentially important and where feedbacks are happening; but, still; the 2D nature of the framework limits it from detecting persistence or recovery; i.e.; the temporal aspect of coupling; especially; when it comes to transitional climates. In general, as pointed out by De Riddler (1997); the potential for convective precipitation increases with evaporative fraction; unless conditions are extremely dry. But, Koster et al. (2004, 2006) has already pointed out that for transitional climates; where the weather is neither dry nor wet; though evaporation is large enough to influence precipitation; the magnitude of convection is still dependent on soil moisture.

So, for the current study; where a 3D framework is considered; it was seen, that, much like the study carried out by Roundy et al (2012); the persistence of the coupling events is controlled by the SM. The initial SM play a key role; until there is a forcing from the atmosphere, which completes the land-atmosphere coupling process and acts as a feedback. Generally, out of the CTP and HI; the initiation of the coupling events is controlled by the humidity index; and the manifestation of such an event is seen in precipitation more than in evaporation. The tendency of SM was seen to be towards an equilibrium; when the moisture content in the land and atmosphere is equal.

For future, I wish to extend this study over a larger spatial area and a larger period. It would also be interesting to study the roles of other variables, like potential evaporation, evapotranspiration, surface fluxes and their impacts on one another over the study area. Since, the study area suffers from an inherent problem of sugarcane plantation (since the late 90's), effect of orography (certain parts of it lie in the rain shadow region of the Western Ghats) and dynamic circulation (ENSO, MJO & monsoons); later works might look up into each of these aspects and form a comprehensive conclusion of the overall effect of all of these factors that might result in drought here.

Application of the index so developed can be manifold. Since the prediction mostly remains to drought to continue; India, being an agrarian economy; can take steps to stop the drought. This can be done by making policies from the government; or, from the normal people. Agricultural practices can be changed there; for example, people can switch from sugarcane (a crop that consumes a lot of water) to barley or wheat (crops that grow on dry soil). This might help in increasing the water content in the soil; which can, in turn, help in drought recovery.

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BIOGRAPHY

Shreya Guha graduated from Gokhale Memorial High School, Kolkata, India, in 2011. She received her Bachelor of Science from University of Calcutta in 2014. She received her Master of Science in Atmospheric Science from University of Calcutta in 2016 and was employed as a junior research fellow in ISRO for six months back in India and as a graduate research assistant for about eight months in George Mason University, Fairfax, US.