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Improving The Palmer Drought Severity Index By Incorporating Snow And Frozen Ground

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IMPROVING THE PALMER DROUGHT SEVERITY INDEX BY INCORPORATING SNOW AND FROZEN GROUND

by

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This thesis, submitted by Shaoyue Qiu in partial fulfillment of the requirements for the Degree of Master of Science from the University of North Dakota, has been read by the Faculty Advisory Committee under whom the work has been done and is hereby approved.

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ABSTRACT

Drought causes extensive damage and affects a significant number of people. To quantify the severity of drought and to better monitor drought, drought indices are necessary. The Palmer Drought Severity Index (PDSI) has its advantages of comprehensive and taking the characteristics of drought into concern so that it was regarded as the milestone in the revolution of drought indices (Heim, 2002). The PDSI has been widely used, tested, and modified since its development in 1965. However, a commonly documented limitation to this index is the lack of consideration for snow and the treatment of ground status. In this study, a simplified snow model is included in the self-calibrated PDSI model, which is run in a monthly time step over the continental United States (CONUS) for the past 32 years (1979 – 2010). In the modified PDSI model, the form of precipitation is based on an air temperature threshold and moisture has been withheld and redistributed into the system based on the accumulation and melt of snow. With the snow processes included, all the model variables (evapotranspiration, soil moisture recharge/loss, surface runoff) decreased in winter and increased in spring melt time and soil moisture remained the same in winter and got recharged in spring, the difference between the soil moisture in the original PDSI model without snow effect and the soil moisture in the PDSI model with snow effect is small. The climatically appropriate precipitation increased more than 200% in northern latitude and mountain

regions in spring melt seasons. The absolute value of moisture departure also decreased in winter and increased in summer, which means the inclusion of snow processes made the moisture departure increase during wet condition and decrease for dry condition (varies more). Inclusion of the snow model also allows the PDSI to better capture spring flooding events, which are caused by snowmelt; monitoring drought events also has been improved due to the changes in duration factor for the modified model with snow processes. Finally, the general moisture conditions as well as the trend of moisture change have been examined using both the original model and the modified model including snow. Both models show similar characteristics over the CONUS for the past three decades. The inclusion of the snow model does not qualitatively change these results, and has little effect on the spatial comparability of the index. Effect of frozen soil has also been examined. In this study, this effect is simply tested by shutting down infiltration when ground is frozen. Due to the time lag between ground status and air temperature (which is the determinant for the calculated snowmelt rate), and also the different spatial resolution for these two data sets, most of the spring snowmelt in western US, especially the northern Rocky Mountain regions didn't infiltrate the soil system and became surface runoff. This loss of moisture caused the soil moisture in those regions decreased from a climate mean 35cm to only ~6 cm, which further made the longitude-averaged annual mean PDSI decreased 0.5 to 0.7 index value.

CHAPTER I

INTRODUCTION

Drought

Drought is defined as “a persistent and abnormal moisture deficiency having adverse impacts on vegetation, animals, or people” by the National Drought Policy Commission and is one of the most complicated and least understood natural hazards. Drought can be categorized into three main types: meteorological drought, which emphasizes the precipitation deficit and evapotranspiration; agricultural drought, which emphasizes the soil moisture deficit and its impact on plants, and hydrological drought, which emphasizes the decrease in river stream flow and reservoir level (Rauber et al., 2008). This study is primary concerned with meteorological drought.

Drought is characterized by its severity, duration, and spatial properties. Severity is the degree of the precipitation deficit or the degree of impacts resultant from the deficit. Based on the definition and character of drought, the duration has several meanings. First, drought is a cumulative result of prolonged water shortage. Second, a region can experience wet and dry spells over shorter periods of time while over the long term, the location could experience the opposite condition to the short term. Therefore, the temporal definition of drought is defined on the interests it impacts. Third, the beginning and end of drought often goes unnoticed. What may appear to be just another heavy rain event may turn out to be the last significant rain for weeks or even months to come; an apparent "droughtbuster" rain is sometimes followed by more abundant rains but other times by a return to dry weather (Rauber et al., 2008).

Drought causes extensive damage and affects a significant number of people (Wilhite, 1993). For example, drought is the number one weather-related cause of death worldwide and ranks second in the weather-related causes of property damage within the United States during the past three decades (Rauber et al., 2008). To better understand and predict this phenomenon, it is critical to characterize drought using a variety of meteorological and hydrological parameters and incorporate them into drought indices.

Drought Indices

Since the first attempt to develop an index of drought, more than 150 drought indices have been developed (Zargar et al., 2011). A number of studies have summarized and compared these indices (Hayes 1999; Heim 2002; Keyantash and Dracup 2002; Zargar 2011; Hayes et al. 2011). Amongst all of the indices, the most common used include the: Standardized Precipitation Index (SPI), Standardized Precipitation Evapotranspiration Index (SPEI), Palmer Drought Severity Index (PDSI), Crop Moisture Index (CMI), Surface Water Supply Index (SWSI), and Drought Monitor.

The SPI is based on the long-term precipitation record for each location. The probability distribution of the precipitation is calculated and then transformed into a normal distribution, such that the mean SPI for the location is zero, then, SPI is defined as the precipitation standard deviation from the mean (McKee et al., 1993). Advantages of the SPI include its simplicity and also different time scales of this index allow it to be applied to longer or shorter types of drought for different purposes. Vicente-Serrano

(2010) improved the SPI to include evapotranspiration into the index, which is known as SPEI.

Instead of merely using the precipitation and evapotranspiration deficit, Palmer (1965) combined the economic supply-demand concept with a hydrological water balance model to calculate the climatological appropriate precipitation and moisture deficiency/excess. Therefore, the moisture departure for each place is in a climatological and historical perspective: the moisture departure is based on both the moisture demand (infiltration to soil, evapotranspiration, surface runoff) and the moisture supply (precipitation), so that not merely the current precipitation anomaly but also the previous moisture condition that has been stored in soil as soil moisture is included. Furthermore, for the ambiguity and difficulty of the beginning and ending of drought, the PDSI also takes this into account.

The PDSI was designed to monitor long-term hydrological trends; it had the limitation of missing short-term droughts. In response to this issue, Palmer (1968) developed another index, the CMI, for agricultural purposes. This index uses the same model and parameters as the PDSI, instead, the CMI is run weekly instead of monthly and less emphasis is placed on previous periods. This allows the CMI to rapidly respond to short-term conditions relevant to agriculture. However, the limitation of the CMI is its incapability of evaluating long-term events. Furthermore, the CMI is only calculated during the growing season, which makes this index incapable of depicting droughts that extend over several years (Hayes 1999).

The SWSI (Doesken et al. 1991) was developed to diminish the limitations of the PDSI in mountainous areas by accounting for snowpack and delays in runoff. This index measures snowpack, precipitation, streamflow, and reservoir storage and calculates the percentile of these terms based on the historical record. Despite improvements in the mountains, SWSI has limits for its application. SWSI is calculated uniquely in each basin and it is difficult to compare SWSI values between basins or regions (Hayes 1999).

The Drought Monitor (DM) is a collaborative effort by agencies within NOAA, the U.S. Department of Agriculture (USDA) and the National Drought Mitigation Center to produce a weekly Drought Monitor product (Heim 2002). As drought monitoring for different purposes have different time scales, emphases and definitions for drought, the Drought Monitor combines a number of indicators and indices: the PDSI, CMI, soil moisture percentile, daily stream flow percentile, SPI, and objective short and long-term drought indicator blends. Besides these five major indicators, the DM uses numerous supplementary indicators for different interests. For example, it also includes the USDA/NASS Topsoil Moisture, Keetch-Byram Drought Index (KBDI), and NOAA/NESDIS satellite vegetation health indices during the growing season. It uses snow water content, river basin precipitation, and the SWSI for snow season. Additional local reports on crops and reservoir status are also taken into account. Based on all the information above, the DM is determined subjectively by humans. The DM draws its strength from the collaborative input at the federal, regional, state and local levels and from the objective synthesis of several indices (Heim 2002). The major deficiencies of DM comes from the subjective processes in the development of this product, the limited time period of DM prohibits studies of historical events, and the available format of the product, which makes quantitative comparisons between this product and other drought indices impossible.

Amongst these drought indices, the PDSI has several key outstanding features. It is one of the few comprehensive drought indices that include all of the parameters in the water cycle by including a simplified hydrology model into the calculation of the index, so that the influences of different variables and different processes on the index and the monitoring of drought can be tested (such as snow process in this study). Furthermore, the PDSI provides an opportunity to place current conditions in historical perspective and it takes the ambiguity and difficulty of deciding the end/beginning of drought into account (Alley 2002). Because of these advantages, PDSI is one of the mostly widely used drought indices in the United States, not only for the scientific community but also by policy makers at the federal level.

Despite the popularity of this index, there are several deficiencies and critiques for the PDSI. These include the limited dataset used to develop the weighting and duration factors (Heddinghaus and Sabol 1991), problems with spatial comparability (i.e. high spatial variability of extreme events; Guttman 1992), the lack of snow or frozen soil processes (Heddinghaus and Sabol 1991; Hayes 1996), and the method used to calculate potential evapotranspiration (Dai 2011). Since the discoveries of these problems, a number of studies have sought to better understand or reduce these shortcomings for the PDSI. Heddinghaus and Sabol (1991) modified the calculation of severity during transitional times while Wells et al. (2004) improved the method to calculate the duration factor and the weighting factor.

Despite the number of studies to test and improve the PDSI, little has been done to include snow and frozen ground processes into the PDSI model and also to investigate the influence of snow process on PDSI values, the probability distribution of PDSI, and its performance on characterizing and monitoring extreme events. While snow and frozen soil processes are irrelevant in warmer climates, they are extremely important at high latitudes or mountainous regions; ground and precipitation status (liquid or frozen) play important roles for both the hydrological cycle and for drought/flood events. Fang and Pomeroy (2007) found that over 80% of annual runoff is derived from snowmelt in Canadian prairies despite snowfall comprising only one-third of the annual precipitation. Hisdal et al. (2006) determined that earlier snowmelt and earlier onset of the growing season has led to longer summer droughts in 60% of river basins across Europe. Kuchment (1999) established that although the probable maximum precipitation rate is usually larger than the probable maximum snowmelt rate, the maximum floods of the medium to large rivers of Russia are of snowmelt origin. Therefore, as the PDSI incorporates a hydrological process model, it is important for it to include snow and frozen soil processes.

Purpose of this study

The purpose of this study is to include the snow process into the PDSI model. The impact of this process on the model will be investigated with a focus on how it changes the monitoring of drought and flood events using PDSI values. This thesis is formatted as follows. In chapter two, relevant background material will be provided including a detailed description of the original PDSI model. This will be followed by a discussion of the procedures and methods for snow modeling. In chapter three, modifications to the PDSI model for the inclusion of snow process will be detailed, along with an overview of the datasets used as input into the model. Chapter four contains results and will be comprised of three sections. The first will detail two extreme events chosen as model examples for the explanation of how PDSI and other model variables change with snow included. Then, climatological characteristics of the modified model as well as the trend of moisture condition in the continental United States during the past three decades will be examined using the two PDSI models as well as precipitation anomaly percentage data. Thirdly, the influence of frozen soil on the model will be tested. Chapter five will be the conclusion for this study.

CHAPTER II

BACKGROUND

The PDSI Model

General Description

As mentioned in the previous section, Palmer (1965) developed an index that combined an economic supply-demand concept with a simplified hydrological model. In this section, this model is described in detail; in a later section, this model will be modified to allow for frozen precipitation. Due to the number of variables and parameters used in this model, the reader can refer to Appendix 1 for a complete list and description of variables and parameters used within this thesis.

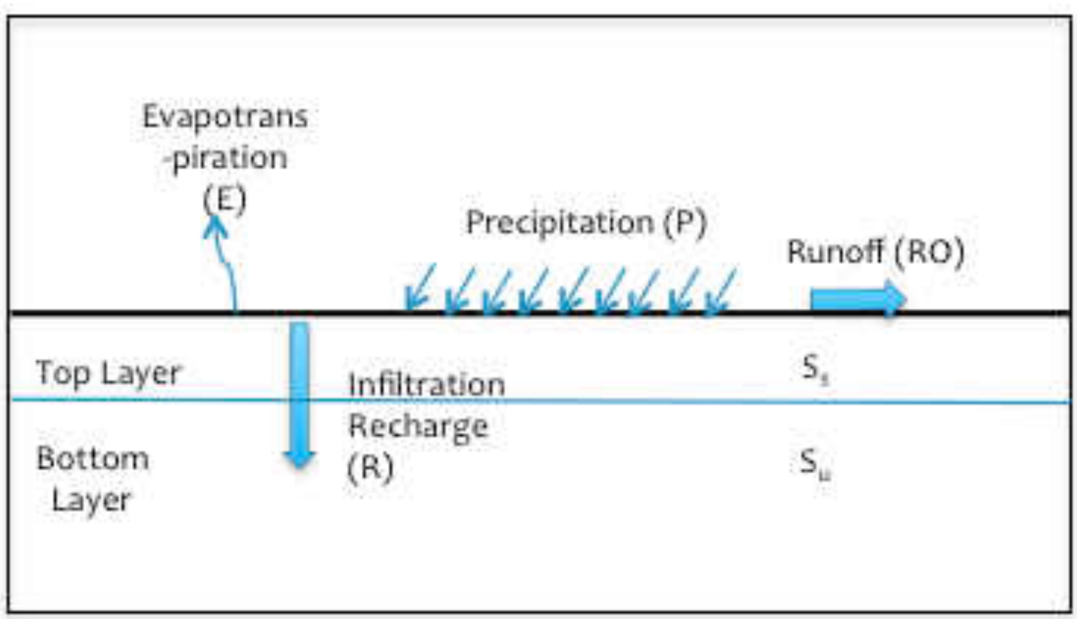

Fig. 1 Two layer hydroloical model for the PDSI.

The foundation for the PDSI is a two-layer soil model with the top layer containing one inch of available moisture at field capacity (S_s) while the available capacity of the soil for the bottom layer depends on soil properties (S_u) (Fig. 1). Moisture supplies the system as precipitation (P) and is lost as evapotranspiration (E) and runoff (RO). To calculate the evapotranspiration, Palmer (1965) introduced a variable known as the potential evapotranspiration ([PE]), which is the amount of moisture that can be used provided the water supply was unlimited. [PE] was originally calculated using the Thornthwaite equation by Palmer (1965) and Wells (2004). It is found that the Thronthwaite method could lead to errors in future climate with rising air tem[rature (Dai et al. 2011). The [PE] calculation has been improved by using the *Penman-Monteith* equation (Eq 1; Dai et al. 2011).

$$[\mathrm{PE}] = \frac{\Delta R_n + \rho_a c_p (\delta e) g_a}{(\Delta + \theta)\lambda_v} \tag{1}$$

Where R_n represents the net irradiance, ρ_a represents the dry air density ($\rho_a = 1.225\ kg\ m^{-3}$), c_p represents the specific heat capacity of air ($c_p = 1005\ J\ kg^{-1}\ K^{-1}$), g_a represents the conductivity of air, θ is the psychrometric constant ($\theta \approx 66Pa\ K^{-1}$), and Δ represents the rate of change of saturation specific humidity with air temperature and is calculated by Eq. 2:

$$\Delta = \frac{de_s}{dT_a} = \frac{5336}{T_a^2} e^{(21.07 - 5336/T_a)}. \tag{2}$$

where T_a is air temperature (K).

δe represents the vapor pressure deficit, or specific humidity and can then be described by Eq. 3:

$$\delta e = (e_s - e_a) = (1 - RH)e_s \tag{3}$$

where RH is the relative humidity, e_s is the saturated vapor pressure of air, e_a is vapor pressure of free flowing air. Finally, the latent heat of vaporization λ_v, is a function of the air temperature (T_a) and is given by Eq. 4:

$$\lambda_v = 2500.8 - 2.361 \cdot T_a. \quad (4)$$

When the net moisture budget is positive (P>[PE]), moisture infiltrates the soil system as soil moisture recharge (R). Once the top layer reaches water capacity, moisture is then allowed to recharge the bottom layer. Conversely, water is removed from the system if all available moisture in the top layer is depleted (i.e. through evapotranspiration). If both of the soil layers reach the water capacity, the remaining moisture leaves the system as runoff ([RO]).

If the net moisture input is negative (P<[PE]), soil moisture loss (L) for the two layers are based on Eqs. 5 and 6:

$$L_s = \min\,[S'_s, ([PE] - P)] \quad (5)$$

$$L_u = ([PE] - P - L_s)\frac{S'_u}{AWC} \quad (6)$$

Evapotranspiration loss from the top layer L_s, is assumed to take place at the potential rate until all available moisture in this layer has been removed; evapotranspiration loss from the bottom layer L_u, depends on the initial moisture content as well as the computed [PE] and the combined available capacity for both soil layers (AWC) of the soil system. Total soil moisture loss is the summation of the loss in both layers: ($L = L_s + L_u$). With this known, evapotranspiration can be calculated by Eq. 7:

$$E = \min\,[[PE], (L + P)]. \quad (7)$$

After the hydrological accounting, to include the climatological information into the index, Palmer also defined potential values for all the terms (recharge, loss, runoff, and evapotranspiration) along with climate mean coefficients. The moisture required for “normal” weather for each month is calculated by the coefficient times the potential terms.

Potential recharge ([PR]) is defined as the amount of moisture that can be added to the system to bring the soil back to field capacity provided it rained sufficiently (Eq. 8).

$$[\mathrm{PR}] = \mathrm{AWC} - \mathrm{S}' \tag{8}$$

Potential loss ([PL]) expresses the measure of a maximum amount of moisture that can be lost from the system provided the precipitation during the period was zero. So, if let P = 0 in Eq. 5, 6:

$$[PL_s] = \min([PE], S_s) \tag{9}$$

$$[PL_u] = ([PE] - [PL_s]) \cdot S_u/AWC \tag{10}$$

The total Potential Loss is the summation of the loss from the surface ($[PL_s]$) and bottom ($[PL_u]$) layers.:

$$[PL] = [PL_s] + [PL_u] \tag{11}$$

Potential runoff ([PRO]) is defined as the potential precipitation minus potential recharge. Palmer (1965) defined potential precipitation as an equal to AWC, such that the [PRO] is calculated as:

$$[PRO] = AWC - [PR] = S' \tag{12}$$

Alley (1984) pointed out that this is an arbitrary approach, however, no better way has been suggested. Thereafter, the four potential values: [PE], [PR], [PRO], [PL] are used to calculated the following climate mean coefficients:

$$\alpha = \overline{\mathrm{E}}/\overline{[\mathrm{PE}]}, \tag{13}$$

$$\beta = \overline{\mathrm{R}}/\overline{[\mathrm{PR}]}, \tag{14}$$

$$\gamma = \overline{[\mathrm{RO}]}/\overline{[\mathrm{PRO}]}, \tag{15}$$

$$\varepsilon = \overline{L}/\overline{[PL]}, \tag{16}$$

The Climatically Appropriate For Existing Condition (CAFEC) for evapotranspiration, recharge, runoff and soil moisture loss are then calculated as the actual potential values times the corresponding coefficients (Eqs. 13-16). The CAFEC precipitation, $\hat{P}$, is then given by the summation of all the CAFEC terms (Eq. 17).

$$\widehat{\mathrm{P}} = \alpha[\mathrm{PE}] + \beta[\mathrm{PR}] + \gamma[\mathrm{PRO}] - \varepsilon[\mathrm{PL}], \tag{17}$$

The precipitation excess/deficiency (d) is then simply the difference between the actual precipitation and the CAFEC precipitation (Eq. 18).

$$d = P - \hat{P}. \tag{18}$$

Calculation of the Weighting Factor

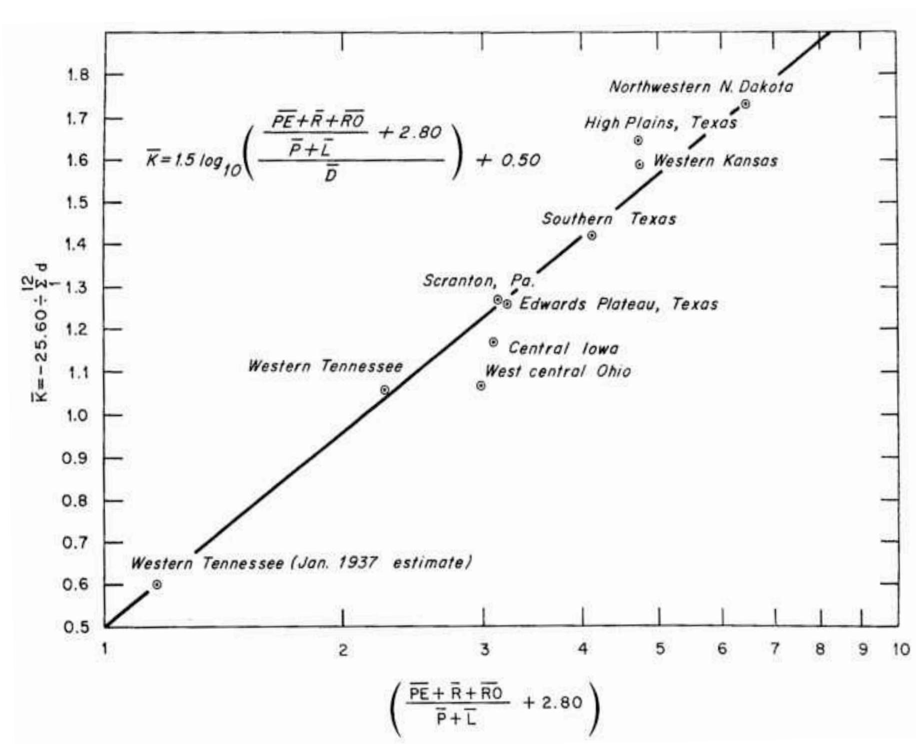

Fig. 2 Relationship of weighting factors to average moisture demand, average moisture supply, and average moisture departure from Palmer's (1965, p. 24 Fig.3).

After the calculation of the *d*, these quantities are standardized over time and space by a weighing factor *K*. The purpose of this weighting factor is to adjust the departure from normal precipitation such that the index values are comparable among different areas and times. For example, a specific index value of precipitation departure during January in the central United States will have the same meaning as the same value in July in the south-east United States even though the precipitation deficiencies will be quite different. To derive this weighting factor, Palmer (1965) used climate data from 9 climate divisions in the US and found a logarithmic relationship between the summation of precipitation excesses/deficiencies and moisture demand, supply and departure (Fig. 2).

The K factor is then calculated as below:

$$K' = 1.5log_{10}[\frac{\overline{PE_J}+\overline{R_J}+\overline{RO_J}}{\overline{P_J}+\overline{L_J}} + 2.8/|\overline{D_J}|] + 0.5 \quad j=1,2, \ldots 12. \tag{19}$$

$$K = \frac{17.67}{\sum_{j=1}^{12}\overline{D_J}K'}K' \tag{20}$$

$$Z = dK \tag{21}$$

where $|\overline{D}_J|$ is the monthly mean of the absolute values of moisture deficiency/excess, *d*. This standardized precipitation excess/deficiency is called *Z* index.

Calculation of the Duration Factor

Based on the character of drought (i.e. the persistence of a precipitation deficit) a duration factor was added into the index. The purpose of this factor is to disallow a single wet (dry) month in a persistent drought (wet spell) to have a great influence on the severity index. To prevent this, weights are given to the PDSI value of the previous month and to the Z index of current month to determine the PDSI (Eq. 22).

$$\mathrm{PDSI_i = pPDSI_{i-1} + qZ_i} \tag{22}$$

To derive these two weights, p and q, Palmer (1965) used the summation of Z indices for the 13 driest events in the two original study areas (central Iowa and western Kansas) and found a linear relationship between the summation of Z index and the length of those droughts as shown in Fig. 3 and Eq. 23.

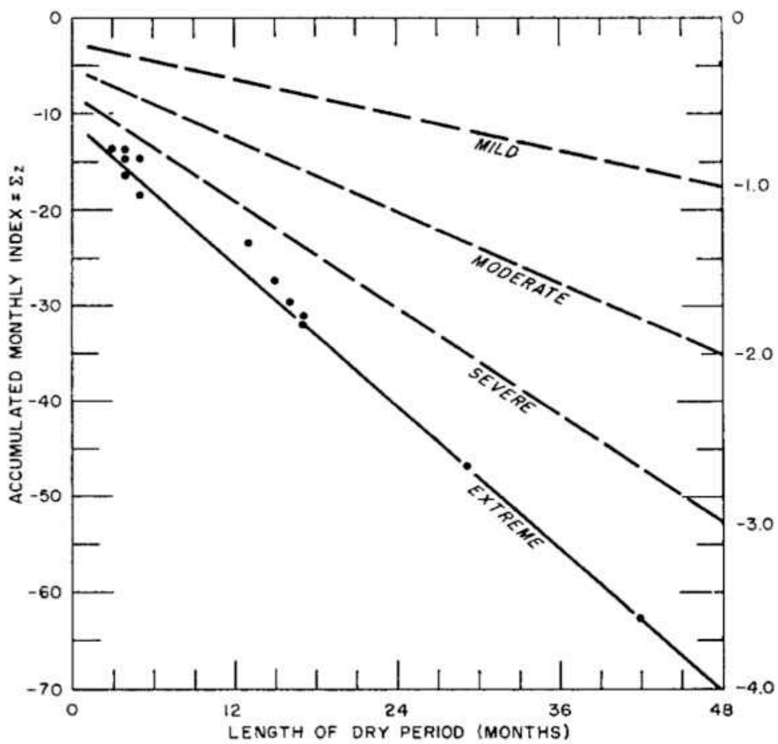

Fig. 3 Accumulated Z index during the 13 driest periods in Palmer's case from Palmer's (1965, p. 20 Fig.1)

$$\sum_{i=1}^{t} Z_i = mt + b \tag{23}$$

With t being the duration of the drought/wet spell in month, p and q are then calculated as:

$$p = \left(1 - \frac{m}{m+b}\right), \tag{24}$$

$$q = \frac{C}{m+b}, \tag{25}$$

where C is the value of calibration index. Palmer (1965) used the most extreme drought cases in western Kansas and central Iowa and scaled the largest value to equal -4.0 (C = -4.0).

Beginning and End of Dry and Wet Periods

To account for the uncertainty for the ending and beginning of drought, Palmer developed a way to determine these properties. The first step of this process is to define the threshold for drought(wet spell); when the severity reaches -1.0 (+1.0), the drought (wet spell) starts, and when the severity reaches -0.50 (+0.50), the drought (wet spell) ends. Then, Palmer calculated three sets of PDSI values X_1, X_2, X_3, using Eq. 22. X_1 is for a wet spell that is "becoming established", so that X_1 is nonnegative, when X_1 is smaller than 0, it is set to 0 again; also, if X_1 is bigger than 1.0, which means a wet spell has "established", this value is given to X_3, and X_1 is set to zero. Similarly, X_2 is for a drought spell that is "becoming established", whenever X_2 is bigger than 0 or smaller than -1.0, it is set to zero. Whenever a dry (wet) spell has established, X_3 is calculated (for details, refer to Chapter 10, Palmer, 1965). During an established dry (wet) spell, if there is a month that the Z index is positive (negative) and tend to end this dry (wet) spell, Palmer calculated the possibility that the dry (wet) spell might end, P_e. If this single month is followed by a continuously dry (wet) condition, the drought (wet spell) will not end, and P_e goes back to zero, PDSI value use the X_3 value. If this single month is followed by more wet (dry) conditions, and finally reduce the drought (wet spell) to normal condition (P_e reaches 100 or the index value drops between +0.5 and -0.5), then the PDSI value used the X_2 or X_1 value. Therefore, the PDSI value of the current month not only depends on the historical moisture conditions but also depends on the future moisture conditions, especially during transitional periods.

Based on the above calculation process for the Beginning and end of dry and wet periods, PDSI suffers some problems at operational level: the PSDI value cannot be assigned until P_e reaches 0 or 100, so that the value may not be known until few months later (Heddinghaus and Sabol 1991). This problem has been fixed by Heddinghaus and Sabol (1991) as well for the calculation in transitional period. This modified PDSI has been widely used hereafter, and sometimes known as PMDI or PDI.

The Self-calibrated PDSI

Because of the deficiencies of PDSI as discussed earlier, Wells et al. (2004) developed a self-calibrated PDSI to improve the calculation of the duration factor and the weighting factor in the PDSI. When deriving the duration factor, Palmer used a limited dataset, arbitrary rules for defining the intensity of the index and an "eye ball" fit method. In Wells et al. (2004), the "eye ball" fit method has been replaced by least square linear regression method; the universal equation has been replaced by the linear relationship for each location using historical data in each location. Also, to solve the arbitrary category problem, they suggested the index can be calibrated to any dry or wet category, but calibrating to extreme wet and dry spells is the easiest to defend for the following two reasons: 1) Palmer calibrated his index to extreme dry spells, and 2) a frequency of 1% - 3% for extreme wet and dry conditions are defined in the literature, whereas there is no commonly accepted frequency of events for Palmer's other wet and dry categories. Finally, Wells suggested calibrating the index value for both extreme wet and extreme dry spells. This is important because some locations may have different sensitivity to wet periods and dry periods.

Similar to the calculation of the duration factor, Palmer only used nine areas' climate data in the United States, mostly from 1931 to 1960, to derive the weighting factor in the model and then applied them over the entire United States for all time periods. In the self-calibrated PDSI, Eq. 20 was replaced by

$$\mathrm{K} = \begin{cases} \mathrm{K}'(-4.0/2\text{nd percentile}), & \text{if } \mathrm{d} \leq 0 \\ \mathrm{K}'(4.0/98\text{th percentile}), & \text{if } \mathrm{d} \geq 0 \end{cases} \tag{26}$$

By doing so, the 2nd and 98th percentile of PDSI values are calibrated to the value of -4.0 and 4.0 for each location. Therefore, PDSI has a clearer meaning for the index value and better spatial comparability over the United States. But Wells also noted that this method is solely based on the moisture departure value d of the location, so that the length of the record as well as the time range of the record has a big influence on the K value. In other words, the PDSI values in 2000 will be different if they are calculated using data from 1975 to 2000 than they would be if calculated using data from 1948 to 2000.

Snow Process Modeling

Snow processes can be classified into two parts: accumulation and ablation. In lieu of direct observations of precipitation type, assumptions must be made for whether precipitation falls in a liquid or frozen phase. The occurrence of snow accumulation is often determined by a temperature threshold (Fontaine et al., 2002; Slater et al., 2000; Pomeroy et al., 1998); when air temperature is ≤ 0° C, precipitation falls as snow and the water equivalent of snowfall is accumulated on top of the ground.

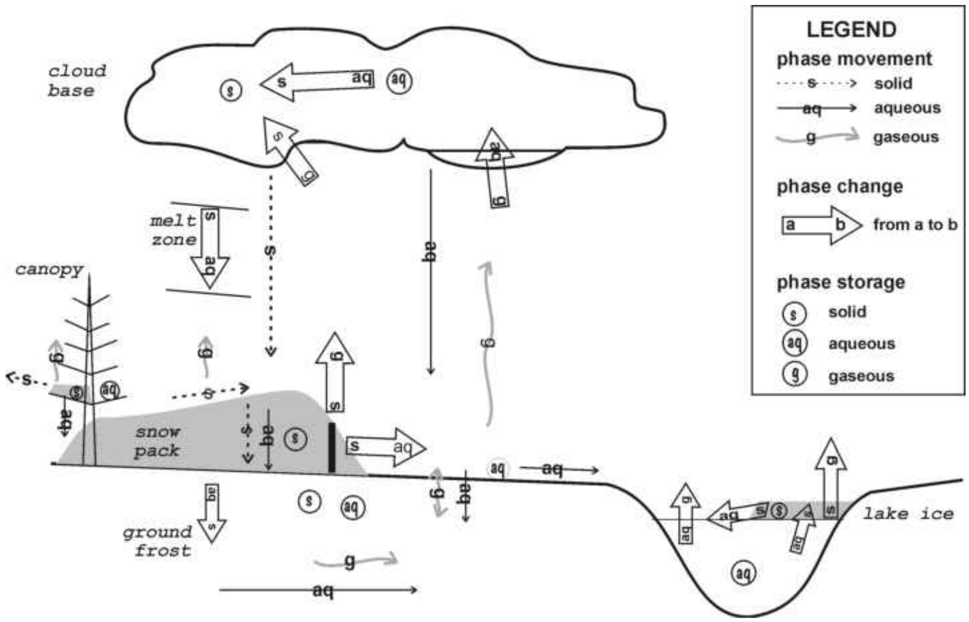

Fig. 4 Water balance and phase change in snow process (from Snow Hydrology Course Notes at Colorado State University)

As shown in Fig. 4, in the snow accumulation process, snow can be redistributed by the vegetation interception and by blowing snow effects. As been found by many studies, these effects can be important for the accumulation process. Pomeroy et al. (1998) conducted research in prairie, boreal forest, and arctic regions in Canada to study these processes. It was found that the amount of interception by the vegetation is a function of the initial snow load on the tree, snowfall speed, canopy density, tree species, LAI and fresh snow density. The influence of blowing snow processes on snow accumulation depends on downwind blowing snow transport and sublimation. These processes, however, are primarily concerned with microscale variability in snow amount. Over larger areas, the positive and negative anomalies tend to cancel out and these processes are often assumed to be negligible. At large spatial scales, such as the size for calculating the PDSI or for a Global Climate model, sublimation is the most important process on blowing snow flux (Pomeroy et al. 1998). Sublimation along with evaporation are primarily a function of the vapor pressure of the air. This process can cause a wide range of water loss for the total snow water equivalent of the snowpack, ranging from 15-40% over Canadian Prairies (Pomeroy et al. 1998), to 5-10% in an Alpine region (Bayard et al. 2005), to 0-30% in a Colorado forest canopy (Montesi et al. 2004).

As illustrate by Fig. 4, Snow ablation process occurs in many forms: evaporation, sublimation (discussed previously), along with traditional snowmelt. Modeling snowmelt can be accomplished using two approaches: with an energy budget or with aid of a temperature index. Although the energy balance method better matches the complex processes of melting snow, the degree-day method that uses a temperature index has been widely used due to its simplicity, the wide availability of temperature data, as well as its good performance for larger spatial and temporal scales (Hock 2003). Furthermore, the WMO (World Meteorological Organization) has established the degree-day method as a standard tool. As a result, this method is common for snowmelt runoff modeling, use by operational hydrologists, and has accuracy comparable to more complex energy budget formulations (Rango and Martinec 1995).

The Degree-day Method for Snowmelt Modeling

The degree-day method calculates the daily snowmelt depth, M, by multiplying the number of degree days, T (°C d), by the degree-day ratio, a (mm °C$^{-1}d^{-1}$) such that:

$$M = a \cdot T. \tag{27}$$

The key component of this method is the determination of the degree-day ratio. Since its development, the degree-day method has been tested, compared with the energy-balance approach, and has been improved. Instead of using one single value of the degree-day ratio throughout the entire snowmelt season, it should vary seasonally and by location (Rango and Martinec 1995; Weiss and Wilson 1958; Bengtsson 1980). WMO has suggested values for this ratio as shown in Table 1 for different vegetation covers and different times of the year.

Table 1 Degree-Day Ratios ($mm\ ℃^{-1} d^{-1}$) as Proposed by the World Meteorological Organization (WMO, 1964)

Month	Moderate Forest Cover	Partial Forest Cover	No Forest
April	2.0	3.0	4.0
May	3.0	4.0	6.0
June	4.0	6.0	7.0

A common problem with these factors was lack of consideration of snow density. Martinec (1960) and Kuusisto (1980) adjust the degree-day-ratio by incorporating this property which varies by vegetation type (Eqs. 36-37).

Forest: $$\alpha\ (\mathrm{mm}\ ℃^{-1}\mathrm{d}^{-1}) = 10.4\frac{\rho s}{\rho w} - 0.7, \quad (28)$$

Open: $$\alpha\ (\mathrm{mm}\ ℃^{-1}\mathrm{d}^{-1}) = 19.6\frac{\rho s}{\rho w} - 2.39, \quad (29)$$

Where ρ_S is the density of snow and ρ_W is the density of water.

When the degree-day method is applied over a watershed or a basin, the actual meltwater volume depends on both the potential melt and the extent of snow coverage. Rango and Martinec (1995) suggested the following equation for the areal degree-day factor:

$$\mathrm{V} = \mathrm{A} \cdot \mathrm{S} \cdot \alpha \cdot \mathrm{T} \cdot 0.001, \quad (30)$$

where V is meltwater volume (m^3), A is the area (m^2), S is the portion of A covered with snow (decimal number), α is the degree-day factor ($mm\ ℃^{-1} d^{-1}$), T is the number of degree days and 0.01 is the conversion of m to mm. In this method, the snow coverage varies with time and is the most important factor.

The degree-day method was further modified to include the influence of snowpack temperature by Fontaine et al. (2002).

$$M = \alpha(\frac{T_{SN}+T_B}{2} - T_m), \tag{31}$$

$$T_{SN2} = T_{SN1}(1-\sigma) + T_B\sigma, \tag{32}$$

where M is amount of melt water released, T_{SN} is the snowpack temperature, T_m is the melt threshold temperature, T_{SN1} is the snowpack temperature for the last time step and T_{SN2} is for the next time step, T_B is the daily mean air temperature, σ is the snowpack temperature lag factor. Instead of one value for each month, as suggested by the WMO, Fontaine et al. (2002) defined the degree-day ration model as a sine function of the day of year:

$$\alpha = \frac{\alpha_{mx}+\alpha_{mn}}{2} + \sin\left[\frac{(\text{day of year})\pi}{366}\right](\frac{\alpha_{mx}-\alpha_{mn}}{2}), \tag{33}$$

where α_{mx} and α_{mn} are the maximum and minimum melt factor assumed to occur on 12 December and 21 June, and are approximately about 6.0 and 2.0 ($mm\,℃^{-1}d^{-1}$) respectively.

CHAPTER II

METHODOLOGY AND DATASETS

Snow Modeling

The essential modification to the PDSI model is inclusion of precipitation falling as snow. A simple snow model is implemented to keep track of the accumulation and snowmelt of snow on the ground surface. Assumptions are made regarding how the liquid equivalent water is stored on top of the soil, input into the soil system, and is lost as runoff.

In this study, the occurrence of snow accumulation is determined by a temperature threshold. When surface air temperature is either at or below freezing (≤ 0 °C), precipitation can fall as snow and the water equivalent of snowfall is accumulated on top of the ground. The snowmelt calculation is based on the improved degree-day method, as recommended by Fontaine et al. (2002), (Eqs 31-33). The primary difference between Fontaine et al. (2002) and this study is the timescale of the model. Our model is run monthly whereas the model in Fontaine et al (2002) was run daily. Therefore, the 15th day of each month is used in Eq. 33 and the monthly mean air temperature is used in Eq. 31 to calculate the monthly mean snowmelt rate. The total snowmelt for this month is the product of the number of days of the month and the mean snowmelt rate. Since snowmelt often occurs at daily or weekly time scales, the validity of this assumption need to be justified. Based on the structure of the model, this approximation may be acceptable given the following reasons:

1) Because the influence of snowmelt rate on the infiltration is ignored, the moisture deficiency/excess depends only on whether the total amount of moisture that enters the system is above or below normal, so that the time of snowmelt is less important.
2) The total amount of accumulated snow at the beginning of the melting season is the same regardless of snowmelt; the melting is only a redistribution of the amount of moisture entering the system.

Modifications to the Moisture Deficiency/Excess

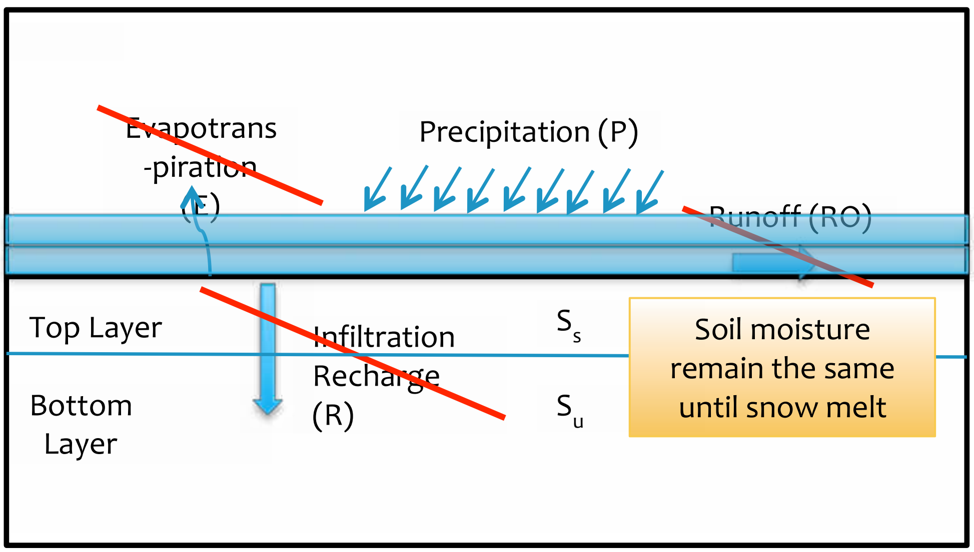

Fig. 5 The hydrological model in winter for the modified PDSI model

The processes discussed above cause moisture to be redistributed over the year. Therefore, both snow accumulation and melting should also be included into the calculation of the moisture deficiency/excess, d, and the CAFEC precipitation, $\hat{P}$. As shown in Fig. 5, when air temperature is below the snow/rain threshold (0 °C), all precipitation accumulates on the ground as snow, evapotranspiration from the ground and vegetation are set to zero (E = 0), and sublimation over snow is also set to zero. When precipitation falls as snow, it is assumed it stays on top of the ground and does not infiltrate into the soil. Therefore, the net moisture input (*P - E*) into the model is zero, soil moisture remains the same, and surface runoff is zero in winter (*L = 0, R = 0, RO = 0*). Because these terms are zero, the coefficients in Eqs. 2-5 are also zero, and thus $\hat{P}$ is zero. Instead, $\hat{P}$ must be redefined during the winter to prevent the PDSI model from potentially providing erroneous information. Under the notion that the long-term mean of $\hat{P}$ is equal to the long-term mean of the actual precipitation (Palmer 1965), $\hat{P}$ for months when air temperature falls below snow/rain threshold is redefined as the climate mean precipitation of the corresponding months. Therefore, the moisture deficiency/excess in winter is simply the precipitation anomaly.

When air temperature rises above the snow/rain threshold, the modified model returns to the original PDSI model as in Fig. 1. Moisture input is then calculated as the snowmelt plus precipitation, and $\hat{P}$ is again calculated by Eq. 17. With the extra moisture source entering the system, the moisture deficiency/excess is redefined as

$$\mathrm{d=P+M-\hat{P}}, \tag{34}$$

where *M* is the amount of snowmelt in that month. It is assumed that all precipitation and snowmelt infiltrate into the ground immediately until water capacity is reached regardless of the ground status. This may be problematic as frozen ground may impede infiltration and partition it to runoff. The effect of frozen ground on the modified model will be further discussed in Chapter 4.

Based on the model setup above, $\hat{P}$ along with the other coefficients will change during months that experience snowmelt. Because these months may vary from year-to-year (i.e. an early or late spring melt), it may be difficult to define the climatological average and moisture deficiency/excess. For example, a location with a climatological melt in March may have years when the snowmelt occurs in February or April or for a location snowmelt usually happen in March and April may have years when snowmelt only happen in March. Early/late melts will cause the *d* parameter in melt months to be abnormally wet while *d* in climatological normal melt months will be abnormally dry even with an average amount of snowmelt for the season.

To resolve this problem, the total amount of snowmelt for the season has been summed up into one month such that the climatological total snowmelt for the season can be defined. This snowmelt month is chosen as the month with the majority of the snowmelt in any given year. Moisture deficiencies in other snowmelt months are defined as the precipitation anomaly. Finally, to account for the variation of snowmelt time over years, only the non-zero values (years with snowmelt) are used in calculation of $\hat{P}$ in Eq. 13-16.

In general, this technique works quite well as snowmelt typically occurs over period of one to two months in length. Other methods were tried such as redistributing the snowmelt evenly or weighting by the climatological average of snowmelt over months. These methods did not work well and frequently led to rapid fluctuations in the *Z* index.

Datasets and Study Area

The original and modified PDSI models were calculated from 1979-2010 for the continental United States. The models utilize precipitation, surface meteorological observation (winds, humidity, temperature), net surface radiation, and soil water-holding capacity data. These datasets are identical to those used in Dai (2011) and are outlined in Table 2.

Table 2 Datasets used in this study

Variable	Data Set	Resolution	Period	Time step	Reference and Sources
Precipitation	GPCP, version 2.2	2.5° × 2.5°	1979-2011	Monthly	Adler, R.F., et al. 2003
Air temperature	CRUTEM3	5° × 5°	1850 - 2011	Monthly	
Surface wind speed, relative humidity, pressure	NCEP Reanalysis	2.5° × 2.5°	1948-2012	Monthly	Kalnay et al. 1996
Surface net radiation	NCEP Reanalysis	Gaussian grid 192 × 94	1948-2012	Daily	Kalnay et al. 1996
Soil water-holding Capacity	Derived	1° × 1°	Climatology	Monthly	Webb et al. 1993

The models were run with a monthly time step on a 2.5° by 2.5° grid, identical to that used by the Global Precipitation Climatology Project (GPCP). All other data were interpolated to the same grid.

CHAPTER IV

RESULTS

Case Study

1. April 1997 Flood in the Red River Basin

Red River Basin (RRB) locates along the border of Minnesota/North Dakota. The annual snow and frozen soil cycle in this area as well as the northward flow of the Red River makes the snowmelt be the main reason for spring flooding. One flood case occurred in April of 1997, which was the most costly flood on a per capita basis for a major metropolitan area in United States history (Todhunter 2001). This flood was caused by a number of factors. The stage was set by substantial precipitation in the fall of 1996 that produced a high level of soil moisture in the basin followed by early freezing of the saturated ground. This prohibited infiltration of additional moisture throughout the winter. These series of events were followed by above normal snowfall in the basin during the winter of 1996-1997 including a series of blizzards late in the winter. Therefore, this flood case is a good example to investigate the importance of including snow effects into the PDSI.

A summary of the PDSI in RRB from March to May 1997 is provided in Fig. 6. The wetness of the period is easily seen regardless of the models with or without snow processes. In the original model, the PDSI values reached the "very wet" (3 to 4) category in April of 1997 (Fig. 6a). In the modified model, which is including the effects of snow, the PDSI values increased the magnitude of the wetness by 0.6 ~ 1 to the "extreme wet" level (Fig. 6b). In addition, the repartitioning of precipitation was also captured in March, when PDSI decreased about 0.2 ~ 1.4 magnitude in the modified model (Fig. 6c).

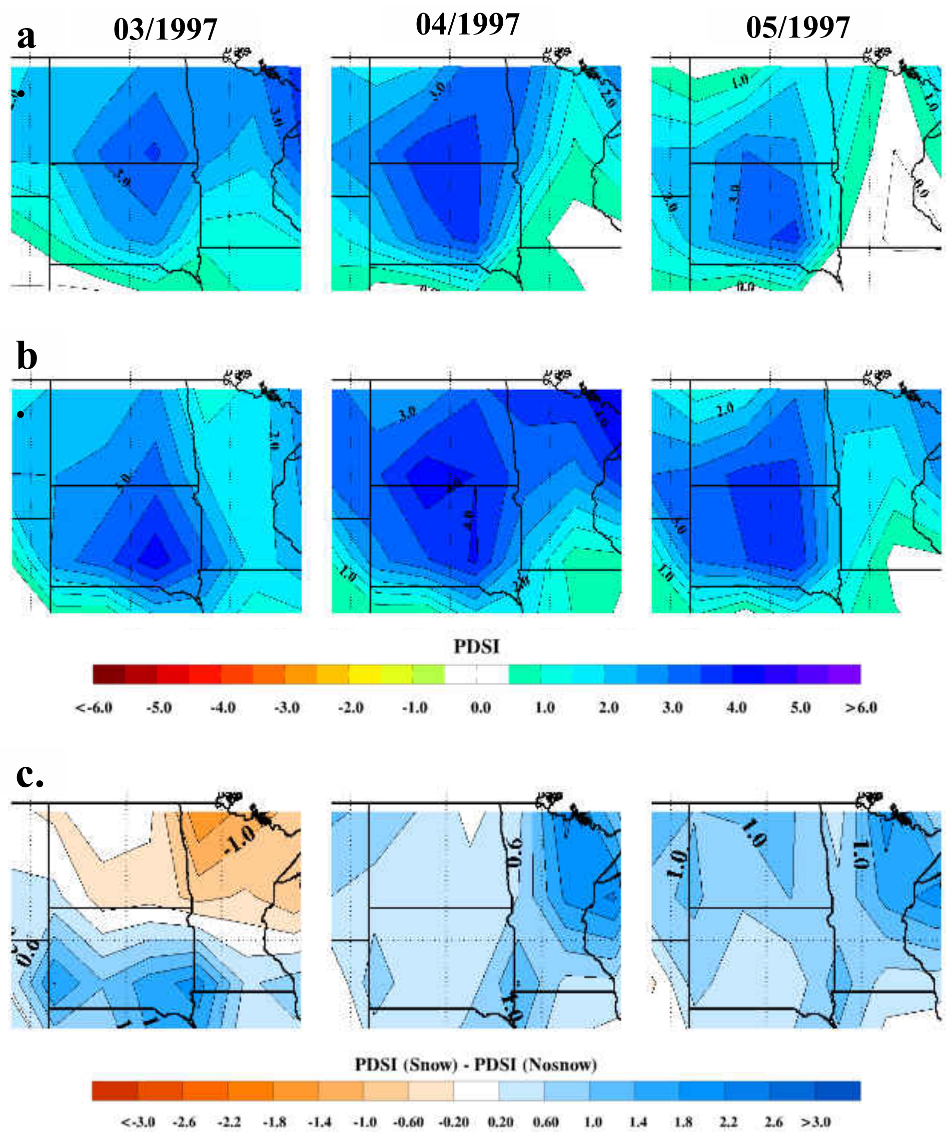

Fig. 6. PDSI contour plot over the Red River Basin (RRB) for (a) original PDSI model, (b) the modified PDSI model and (c) their differences (b-a) for March (left col.), April (middle col.) and May (right col.) in 1997.

To better understand the detailed differences between the modified model and the original model, a time series for environmental and key variables in the model are shown in Fig. 7 for one location of the RRB (96.25° W, 46.25° N) from January 1996 to December 1998.

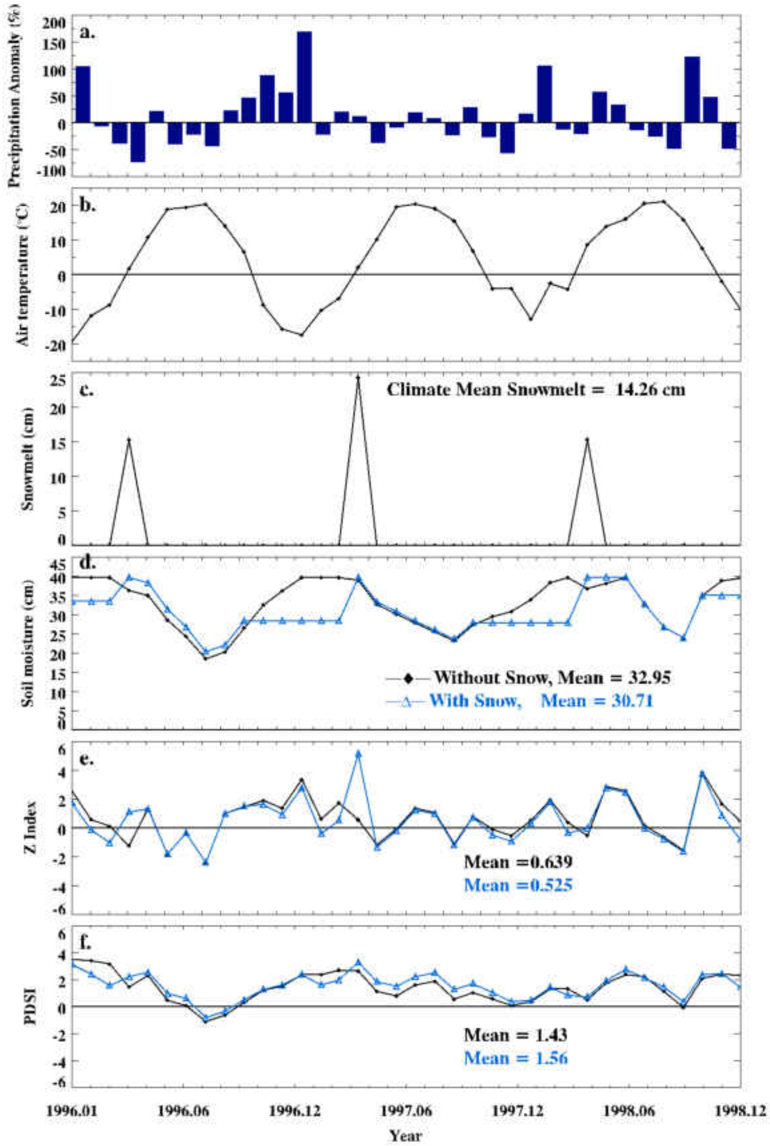

Fig. 7. Monthly means for (a) precipitation anomaly percentage (relative to corresponding averages for the period 1979-2010), (b) air temperature, (c) snowmelt, (d) soil moisture, (e) Z index, and (f) PDSI for the two PDSI models for a point (96.25° W, 46.25° N) over the RRB from January 1996 to December 1998.

In the typical continental climate zone, this location has a regular snow cover cycle and the majority of snowmelt occurs in April (Fig. 7c). Based on the discussion for the modified model, soil moisture remained the same when the monthly average temperature was below zero Celsius and increased with the spring thaw (Fig. 7d). The above normal precipitation from September 1996 to January 1997 (Fig. 7a) resulted in an anomalously high snowmelt in April 1997, and thus a large Z index despite the near-normal precipitation in April of 1997. It must be noted that the Z index in the winters was lower in the modified model. This is because that under wet conditions, $\hat{P}$ will be smaller than the climate mean precipitation (less demand). Since the wintertime $\hat{P}$ in the modified model is replaced by the climate mean precipitation, it is larger than the $\hat{P}$ in the original model, which causes the Z index to be smaller. Even though the Z index for April 1997 was over 5 in the modified model, the duration and weighting factors cause the increment in PDSI to only be ~1.3. This further demonstrated that the PDSI is designed for a long-term impact, such as drought, and is not as sensitive to short-term and intense events, such as spring floods. Despite the relatively small increment of the PDSI, the duration factor made the impact of this spring flood last almost a year.

2. Colorado Drought of 2002

The second case chosen is the severe drought in Colorado during 2002 to 2004. Colorado has its unique climatology due to its inland continental location in the middle latitudes, and its mountains and ranges extending north and south approximately through the middle of the State. The western part of the state is mountainous and rugged with snow cover almost half of the year and snow accounts for ~30% of annual total precipitation; while the eastern part of the state is high plains with much less snow. Therefore, Colorado is a typical example to study the snow effect due to high elevation, especially in lower latitudes.

The 2002 drought in Colorado is the driest individual year in recorded history, with precipitation total reaching lowest on record for many long-term weather stations, and the streamflow volumes only 25% of average (Doesken and Gillespie, 2004). Fig. 8 shows how the two PDSI capture this even as well as the difference in PDSI for the two models from January 2003 to December 2003. In general, with snow effects included, western (northern) CO is dryer (wetter).

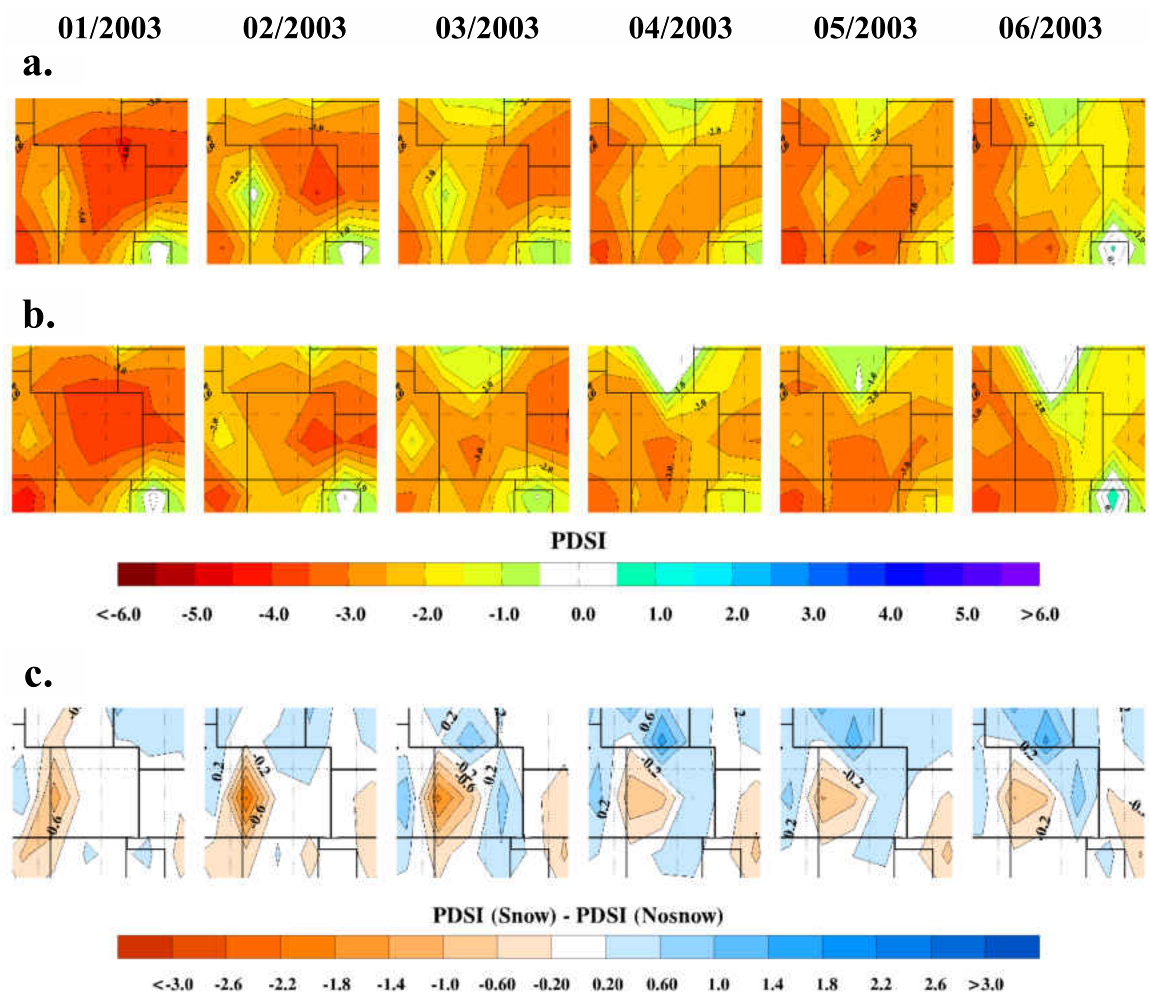

Fig. 8. Same as Fig. 6, except for over Colorado from January to June 2003.

These changes in the PDSI can by explained by looking at time series of key variables as was done in Fig. 7 for grid points in western Colorado (Fig. 9a-f) and northern Colorado (Fig. 9g-l). During early 2002, precipitation anomaly percentages (hereafter, PAP) were ~ -100% resulting in the soil moisture, *Z* index, and PDSI all being indicative of drought conditions.

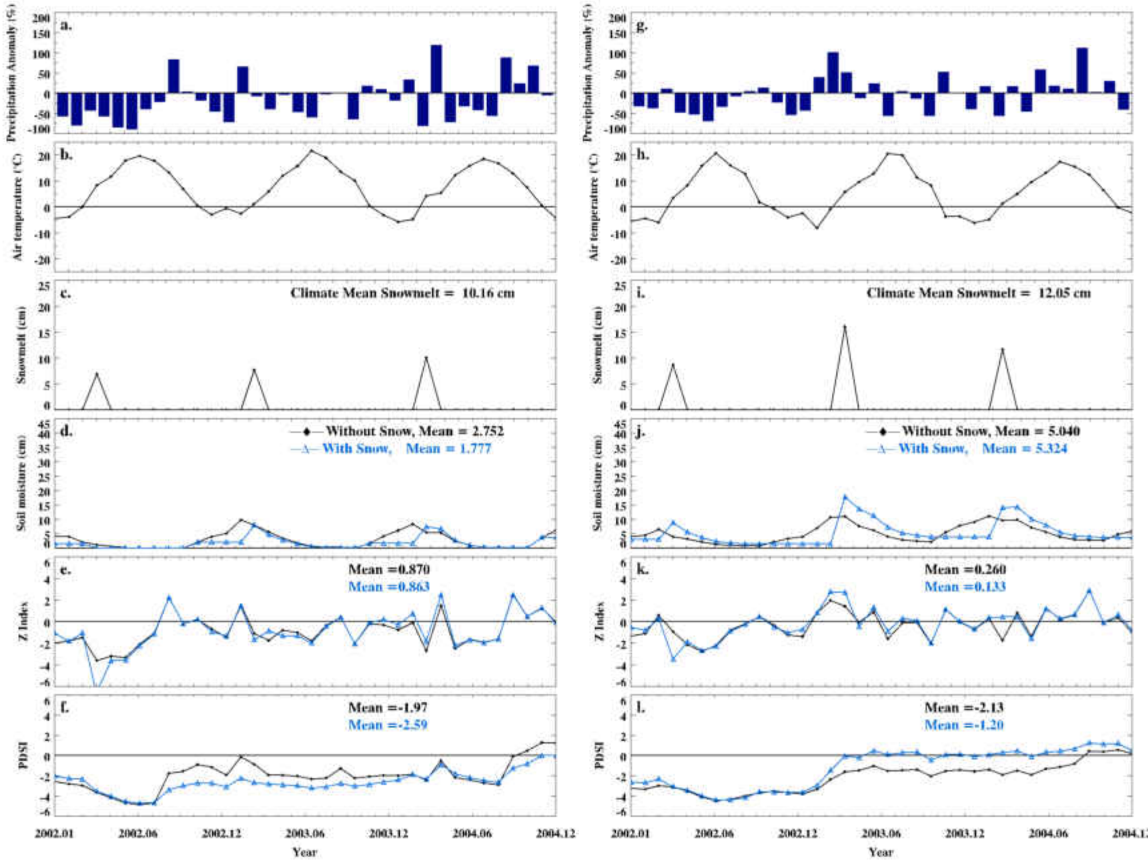

Fig. 9. Same as Fig. 7, except for one surface site over the RRB (108.75° W, 38.75° N) (a) – (f), and over Colorado (106.25° W, 41.25° N) (g) – (l) from January 2002 to December 2004.

In the west (Fig. 9a-f), because of the dry winter from 2002 to 2004, the *Z* index was slightly drier during the spring melting season in the modified model for all the three years due to the less than normal snowmelt. The inclusion of snow effects also changed the *p* and *q* values for the duration factor in Eq. 20: the *q* value for wet cases decreased from 0.55 to 0.3 in western Colorado (this will be discussed in more detail later), so that the wet month in September of 2002 in the western part of the state has less weight in the modified model than in the original one. Therefore, this single wet month did little to improve the drought in the modified model; while in the original model, it improved the drought to the “mild drought” level and even to neutral conditions in January 2003. In reality, this drought remained at an exceptional level, especially in western Colorado, despite the more humid September (Pielke et al. 2005). Therefore, the original PDSI model recovered too quickly from this drought.

The drought in the northern portion of Colorado was reduced by several snowstorms in the spring of 2003 (Fig. 9g-l). Followed by a cool and wet summer east of the mountains, these conditions finally brought the drought to a “moderate” level in May and below drought criteria in July (Pielke, et al. 2005). In the modified model, northern Colorado had above normal snowmelt (15 cm) for the spring of 2003 (Fig. 9i). The *Z* Index was higher for March and April 2003, resulting in a wetter PDSI than in the original model, better representing the observed conditions in northern Colorado.

Climatological Characteristics of the Modified Model

The case studies in the last section demonstrated how key variables in the model are affected by the inclusion of snow processes. At least in these cases, the better physical representation of the water cycle improved the description of the extreme events. What remains to be answered, however, is how these variables change in a climatic sense over the 30-yr duration of this study.

Climatology of Snow

The PDSI model is only altered for regions that receive snow over a duration of time such that the monthly temperature falls below 0° C. To understand where, when, and how often this occurs, Fig. 10 is presented. Snow accumulation and snowmelt occurs in the higher northern latitudes and at higher elevations in the Rocky Mountains with the maxima occurring over the northeastern United States and southeastern Canada (Fig. 10a). With respect to the model, snow water equivalent is responsible for up to 30-40% of the annual precipitation budget in these regions (Fig. 10b). At more southern latitudes or near the coasts, there are a number of grid boxes that only experience snow accumulation and melt for ~50% of the years (Fig. 10c). When averaging the number of months with snowmelt per year, it can be seen that for areas with the most snowfall, the melt frequently occurs over a period extending across 1-2 months (Fig. 10d).

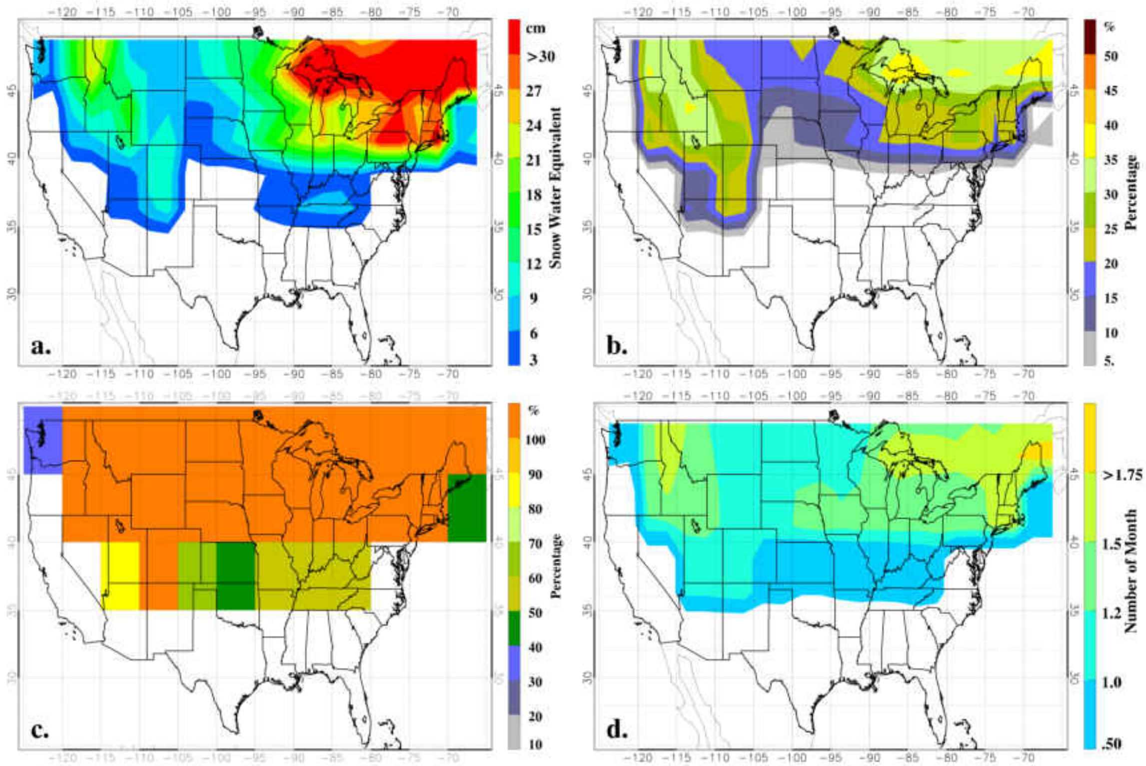

Fig. 10. Climatic means of (a) annual total snowfall (snow water equivalent), (b) the percentage of annual precipitation occurring as snow, (c) the percentage of years with snowfall and snowmelt and (d) the average number of months per year with snowmelt during the period 1979-2010.

Climatology of Model Variables

With the notion that snow occurred 100% of the time in higher latitudes and higher elevations, and accounts for ~30% of the water budget in these regions, the redistribution of water alters a number of model variables. In this section, how the climatological characteristics of all the variables, as well as changes in the snow processes brought to the model variables in different seasons will be investigated. All the variables are shown for the months of January and April, since the model has the most variability during the winter and spring seasons.

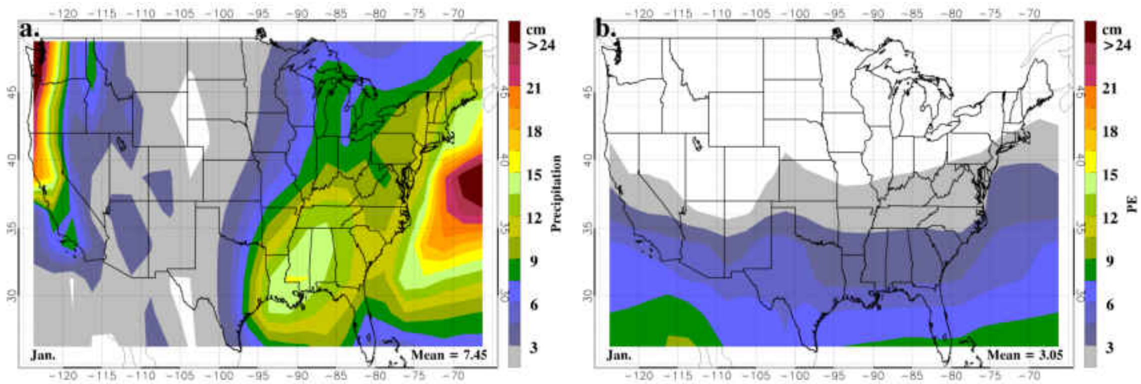

Fig. 11 Climate mean (a) precipitation and (b) calculated potential evapotranspiration in Continental US in January.

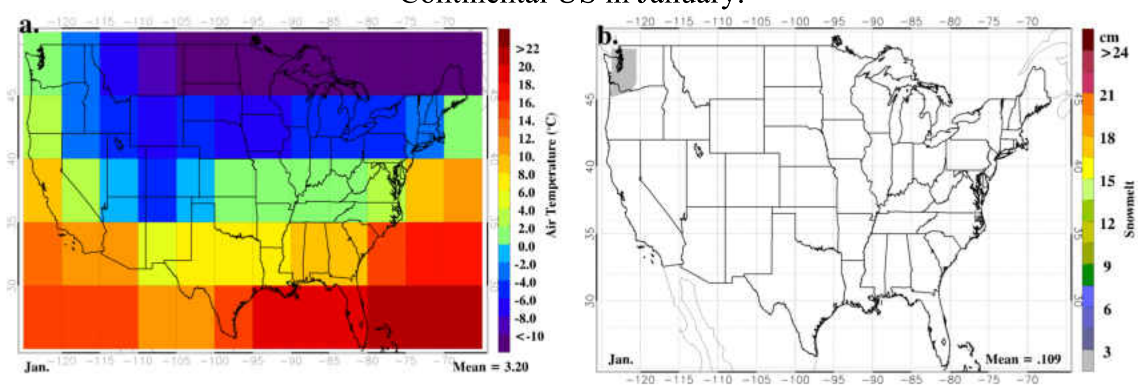

Fig. 12 Climate mean (1979-2010) of (a) air temperature and (b) snowmelt in January

In winter, the west coast and east of United States received abundant precipitation (Fig. 11a) while the potential evapotranspiration is low (Fig. 11b). Therefore, in the original model, most of the precipitation became recharge to the soil and the rest became runoff in west coast and east us (Fig. 13d, g). However, as shown in Fig. 12a, in winter, most of these regions have an average air temperature below 0° C and the ground should be covered by snow and no surface runoff or soil moisture recharge should happen. While in the modified model, all these problems have been fixed: evaporation, recharge, and runoff are all zero in northern latitudes, except for the west coast regions (Fig. 13b, e, h, k).

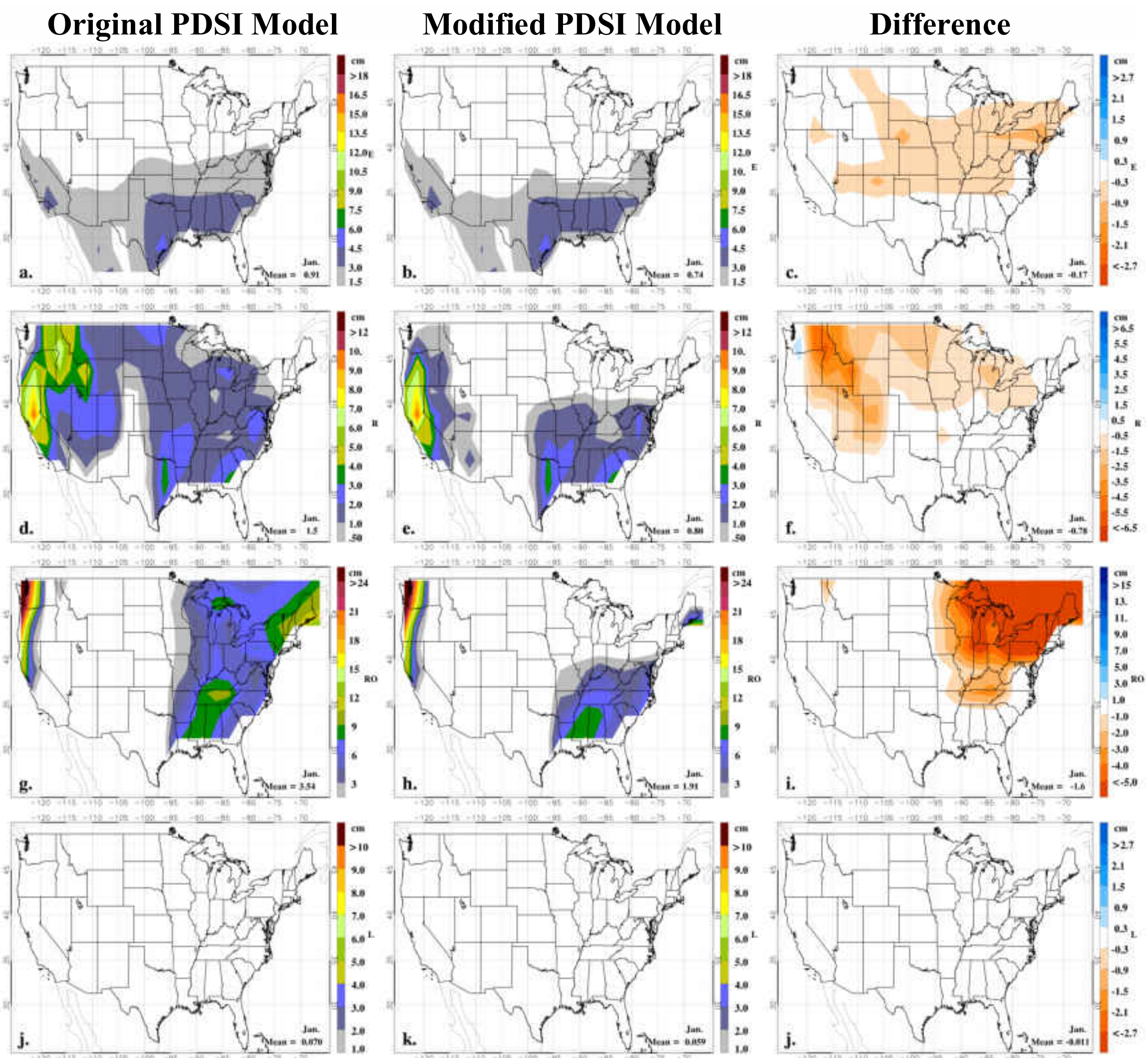

Fig. 13 Climate mean (1979-2010) of (a-c) Evapotranspiration, (d-f) soil moisture recharge, (g-i) runoff, and (j-l) soil moisture lost for the original PDSI model (left col.), the modified PDSI model (middle col.) and the difference (modified-original) (right col.) in January.

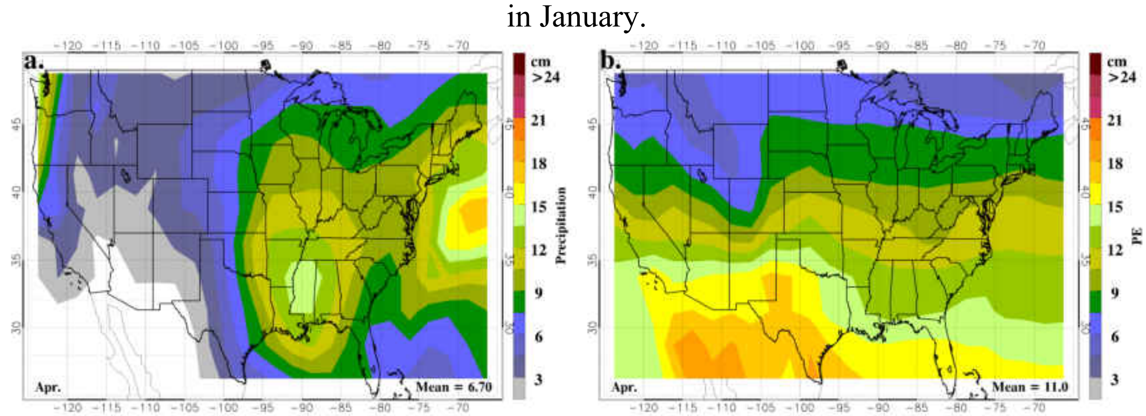

Fig. 14 Same as in Fig. 11, but for April

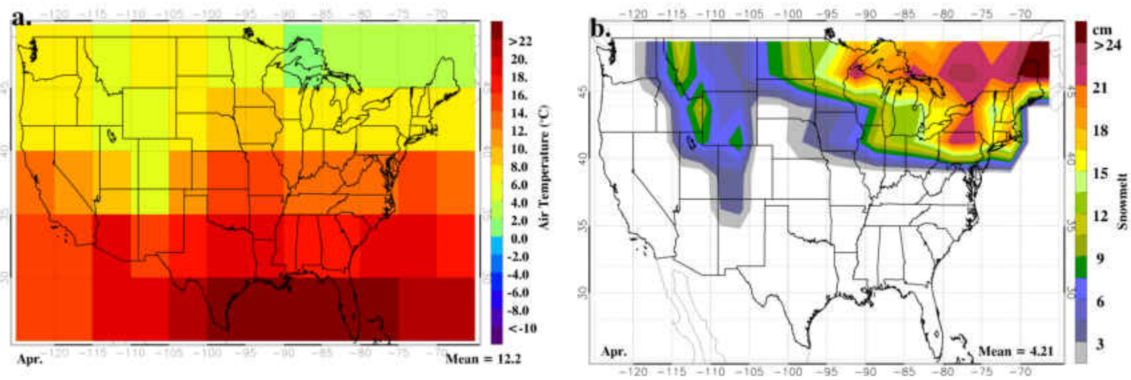

Fig. 15 Same as in Fig. 12, but for April

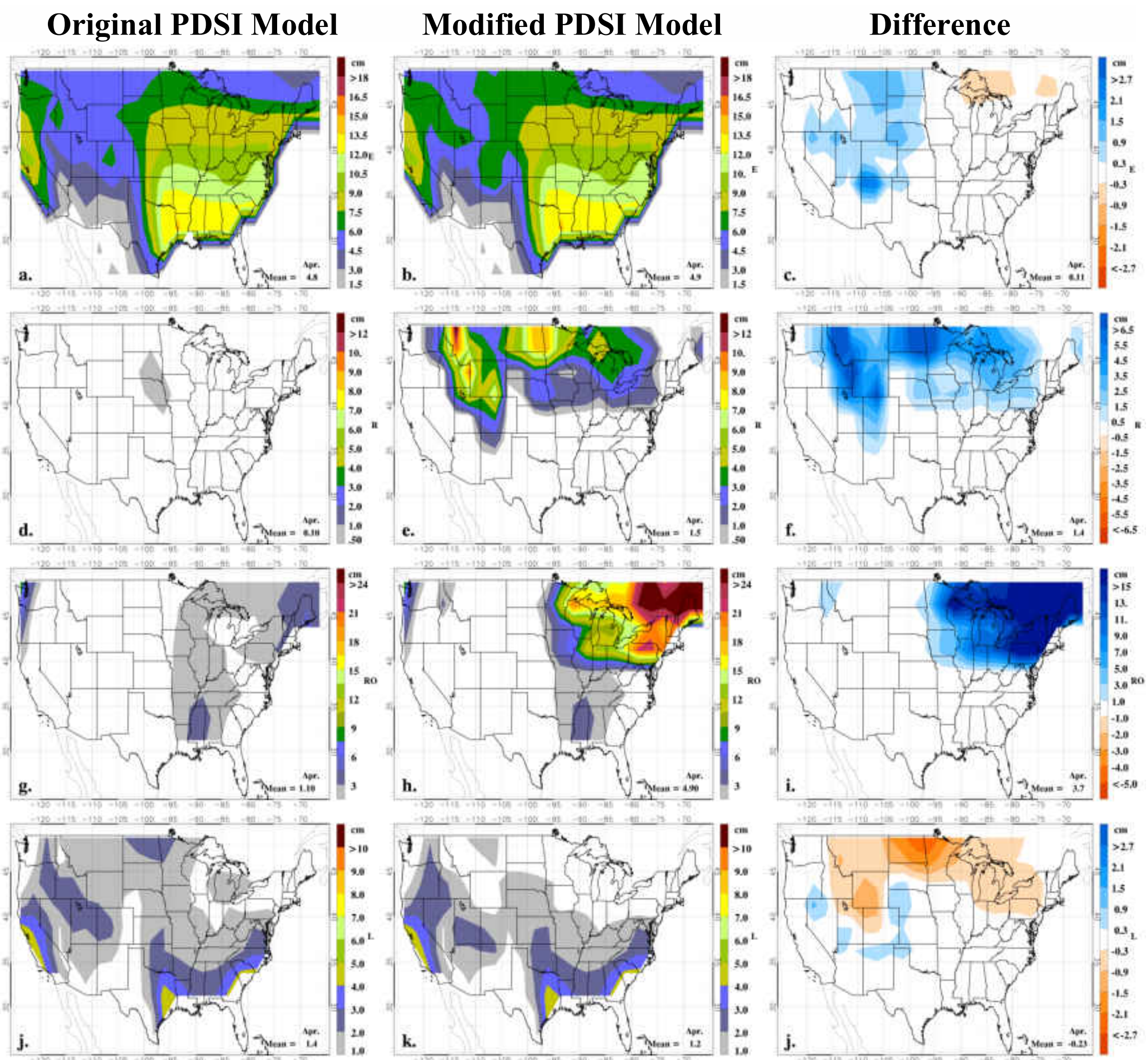

Fig. 16 Same as in Fig. 13, but for April

In the spring melting time, evapotranspiration increased while precipitation did not increase much (Fig. 14a, b), therefore, in the original model, recharge and runoff decreased to almost zero (Fig. 16d, g). With snow effect included, and March and April being the main melting seasons, the released snowmelt not only increased the soil moisture recharge and runoff (Fig. 16e, h), but also increased the evapotranspiration term due to the increase in soil moisture (Fig. 16b, c). As the western US is dry climate and the eastern US is moist climate, all the snowmelt becomes recharge instead of runoff in the west US while most becomes runoff in the east US.

To understand the seasonal variation for all the terms in the model, as well as the influence of snow on these variables, two regions are used as examples to show the seasonal variation. One region is in Rocky Mountain area (105° W - 110° W, 35° N - 42.5° N) (west Colorado, south Wyoming and north New Mexico), as shown in Fig. 17; the other region is in northeast US and southeast Canada (80° W - 85° W, 42.5° N - 50° N), as shown in Fig. 18.

In the Rocky Mountain region, the evapotranspiration has its peak in spring and with the lowest in winter (Fig. 17a). Along with the evapotranspiration, precipitation in this region (black in Fig. 17a) almost evenly distributes over the year, with the lowest value in summer and highest value in autumn. In the original model, due to the annual distribution of precipitation and evapotranspiration, winter is the main season for soil moisture to recharge; spring and summer are the main season soil moisture lost (orange in Fig. 17b,d). As soil moisture is mostly below field capacity (Fig. 17e), the surface runoff in the Rocky Mountain area is almost zero (Fig. 17c)

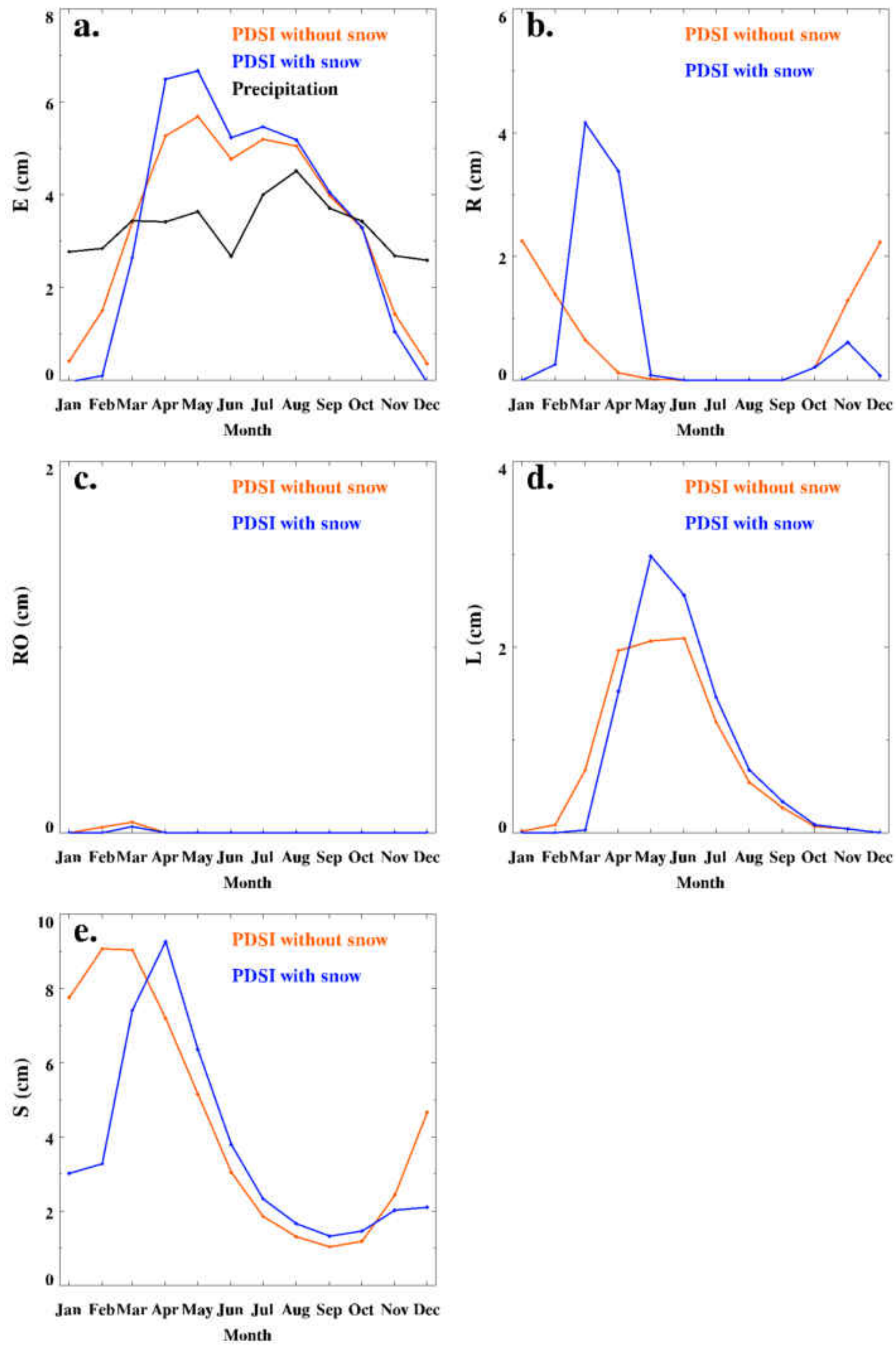

Fig. 17 Seasonal variation of the four model variables: (a) evaporation, (b) recharge, (c) runoff and (d) loss in Rocky Mountain area for the original model (orange lines) and the modified model (blue lines).

In the modified model, the snow effect makes the climate mean evapotranspiration decrease in winter (November to March) and increase from April to

September. This increase in evapotranspiration is due to the change in soil moisture. As in Fig. 17e, soil moisture remains the same in winter (November – February), then it increase in spring melt time and exceeds the original model in April due to the lack of evaporation in winter. The soil moisture recharge also changed its seasonal distribution with snow effect included (blue line in Fig. 17b): R is zero in winter, and then reaches the maximum in spring and return to zero again in summer. In contrast with recharge, the snow effect increased the soil moisture loss in summer due to the increase in soil moisture (Fig. 17d, Eq. 5,6). For runoff, as the climate mean precipitation rate is smaller than the evapotranspiration rate in summer in this region, the soil is always at unsaturated condition, so that the climate mean runoff is almost zero for all the months (Fig. 17c).

Similar analysis is done for the northeast US and southeast Canada region, another region snow takes up to 30% in the annual total precipitation and the climate there (humid continental) is totally different than the Rocky Mountain region. As shown in Fig. 18a, the evapotranspiration still has its maximum in summer and minimum in winter. Different from the Rocky Mountain region, in this region, soil moisture gets recharged in autumn and reaches its water capacity (Fig. 18e), so that most of the precipitation in winter and spring become runoff in the original model (Fig. 18b, c). With snow effect, soil moisture remains the same in winter and doesn't reach the capacity (Fig. 18e), so that the soil moisture recharge has a second pick in spring in the modified model (Fig. 18b). The rest of snowmelt in spring becomes runoff (Fig. 18c).

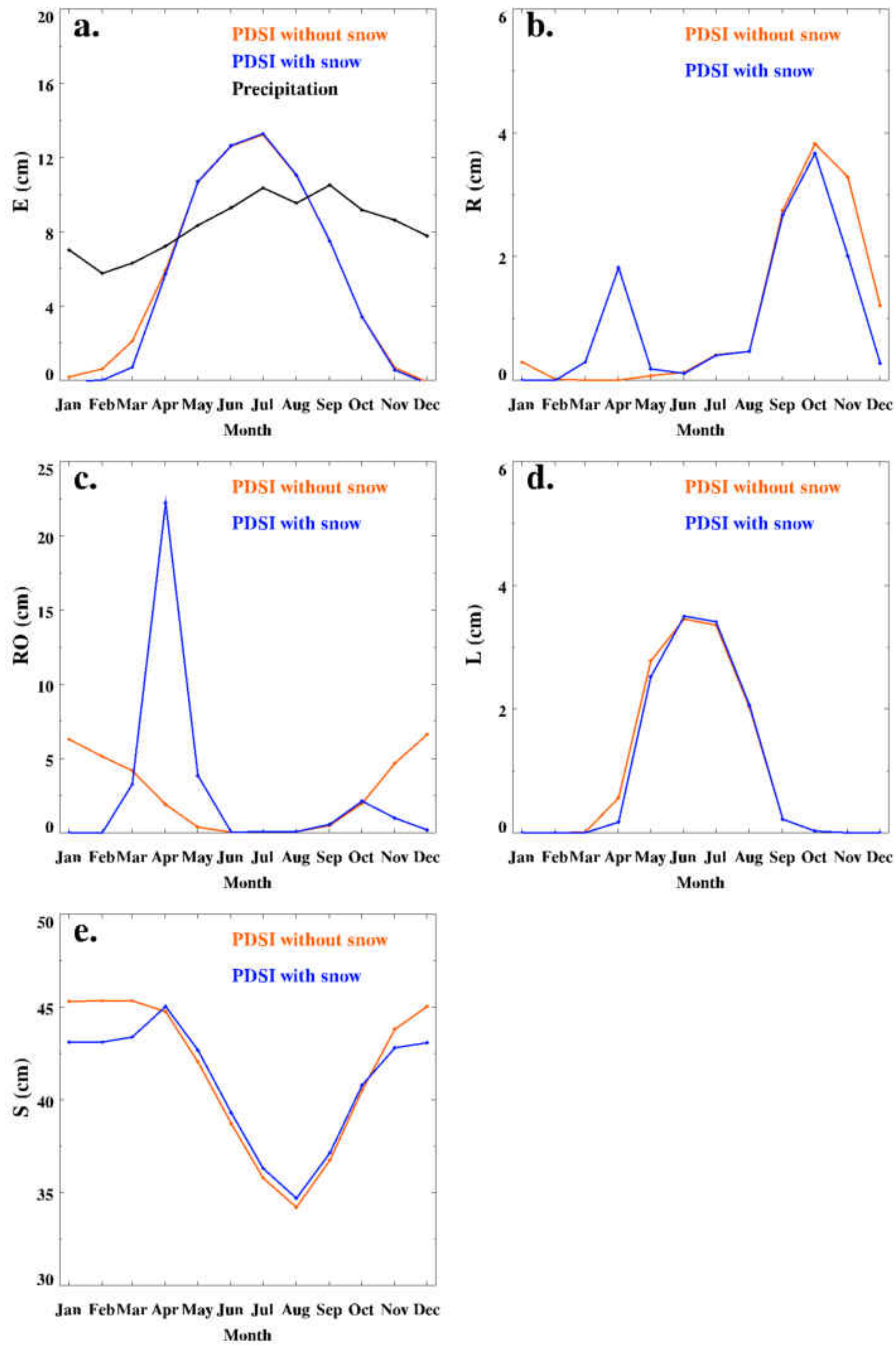

Fig. 18. Same as in Fig. 17, but for northeastern US and southeastern Canada.

With all the changes in the model variables, the key variables in the model: $\hat{P}$ and d also changed. During winter, $\hat{P}$ is consistent with the 32-yr mean precipitation in January (Fig. 11a) with higher values along the coasts and a minimum in the plains (Fig.

19a). It is obvious that with snow effect included, the difference between the two models for $\hat{P}$ is minimal (Fig. 19c). This is because $\hat{P}$ is defined as the climate mean precipitation in the modified model in winter. In a balanced water budgets, the long-term mean of $\hat{P}$ should be equal to the long-term mean precipitation. The only change of $\hat{P}$ in winter is along the west coast due to the early snowmelt in January in that location (Fig. 12b, Fig. 19c).

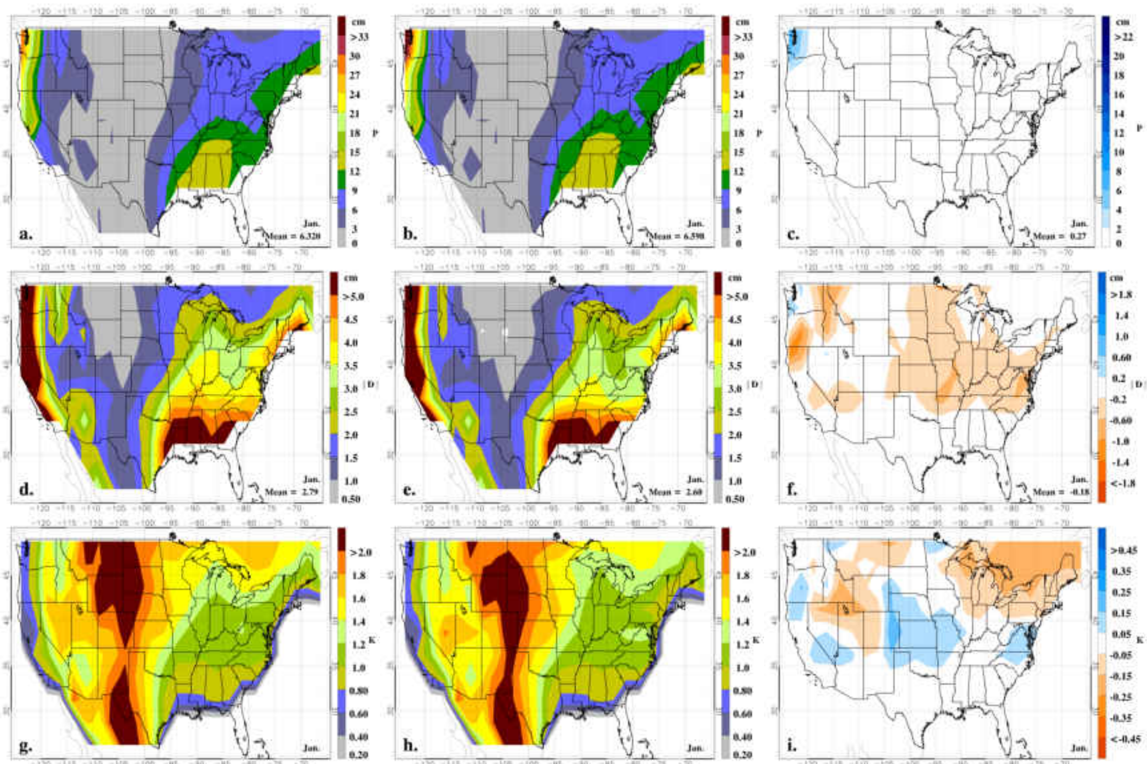

Fig. 19. 32-yr mean (1979-2010) of (a) – (c) CAFEC precipitation, (d) – (f) absolute value Z index and (g) – (i) the weighting factor K, for original PDSI model (left col.), PDSI model with snow (middle col.) and their difference (right col.) in January.

For the absolute value of deficiency/excess, $|\overline{D}|$, and the weighting factor K', $|\overline{D}|$ for both models have similar pattern in CONUS as climate mean precipitation and $\hat{P}$, with higher variation in the west coast and east US and lower variation in the plains (Fig. 19d, e). The weighting factor, K, which is compensate for the spatial difference of d, has the opposite pattern of d; it has higher values in the plains (1.5->2) and lower values in the coast regions (0.5-1) (Fig. 19g, h). With snow effect included, since winter $\hat{P}$ is replaced by climate mean precipitation in the modified model, variation in the moisture departure, *d*, decreased in winter in the modified model (Fig. 19e, f). For the weighting factor K', as R, RO and L are all zero in winter, Eq. 19 become

$$K' = 1.5log_{10}[\frac{\overline{PE_J}}{\overline{P_J}} + 2.8/|\overline{D_J}|] + 0.5 \quad j=1,2, \ldots 12. \tag{35}$$

K' in the modified model decreased in regions where climate mean air temperature drop below 0° C in winter (Fig. 19h, i). For regions that experienced some years that the air temperature drop below 0° C, as $|\overline{D}|$ decreased, *K'* increased.

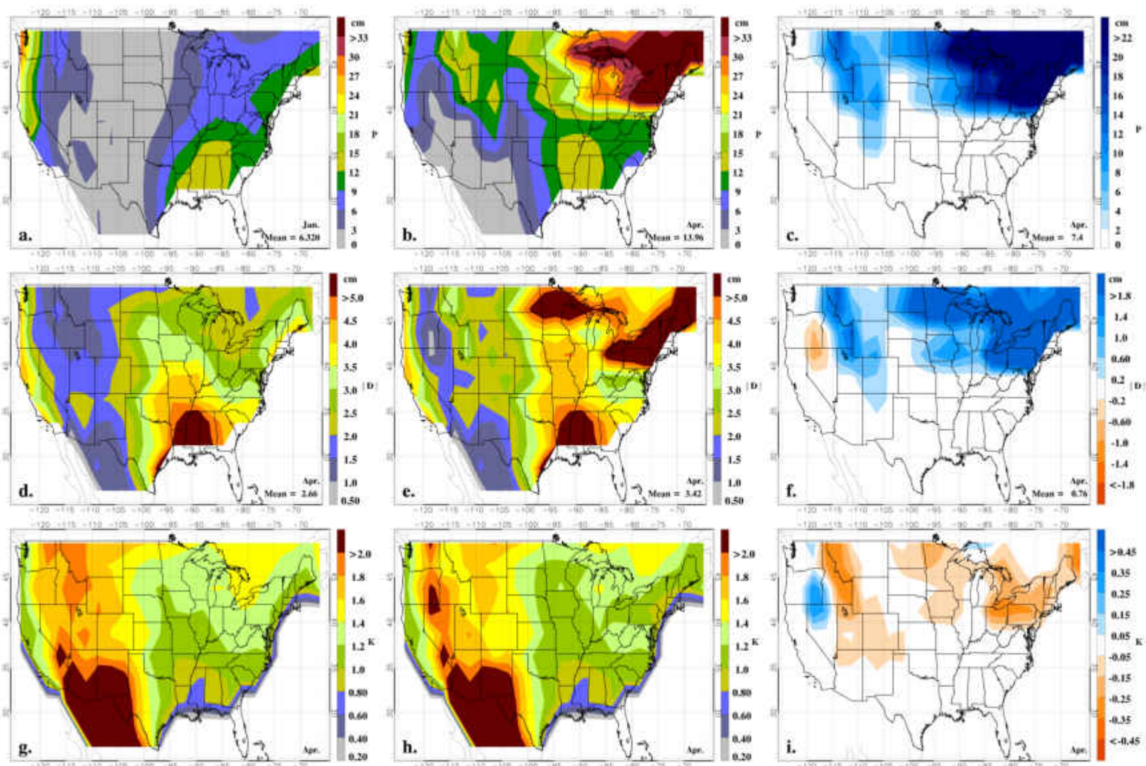

Fig. 20. Same as in Fig. 19, but for April

In spring melting season, $\hat{P}$ increases by more than 200% in the northern United States and the Rocky Mountains due to moisture entering the system from snowmelt. In fact, correlations between the increases in climate mean $\hat{P}$ (Fig. 20c) and the climate mean snowmelts (Fig. 15b) are 0.98. This further demonstrates that it is necessary to include snowmelt as moisture input when calculating the moisture deficiency/excess (Eq. 33), otherwise, the moisture deficiency will always be negative during snowmelt season. The fluctuation of $|\overline{D}|$ increased in spring in the modified model in snowmelt regions, similar as discussed in the case study section, the inclusion of snow made the wet spell wetter and dry spell dryer. In the case of weighting factor K', it still changed inversely with the $|\overline{D}|$.

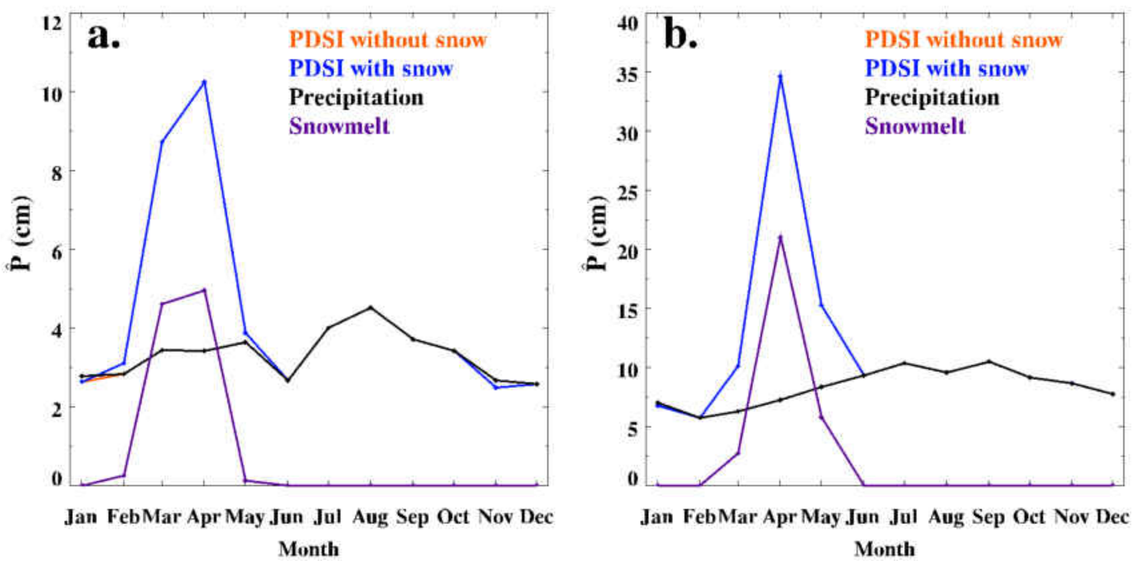

Fig. 21 Seasonal variation for $\hat{P}$ for both PDSI models, precipitation (black line) and snowmelt (purple line) for (a) Rocky Mountain area and (b) northeastern US and southeastern Canada

Seasonal variations for CAFEC precipitation in the previous two study regions are shown in Fig. 21. As discussed before, the long-term mean of $\hat{P}$ is equal to the long-term mean of precipitation (orange line and black line in Fig. 21). In Rocky Mountain region, snowmelt start from February and the main melting season is March and April with climate mean snowmelt around 5 cm. The climate mean $\hat{P}$ in the modified model increased from ~4 cm to ~10 cm in melting season. The climate mean increment of $\hat{P}$ is more than the climate mean of snowmelt. Pattern of $\hat{P}$ as well as the changes made by snow effect is similar in the northeastern US and southeastern Canada region, except for larger values.

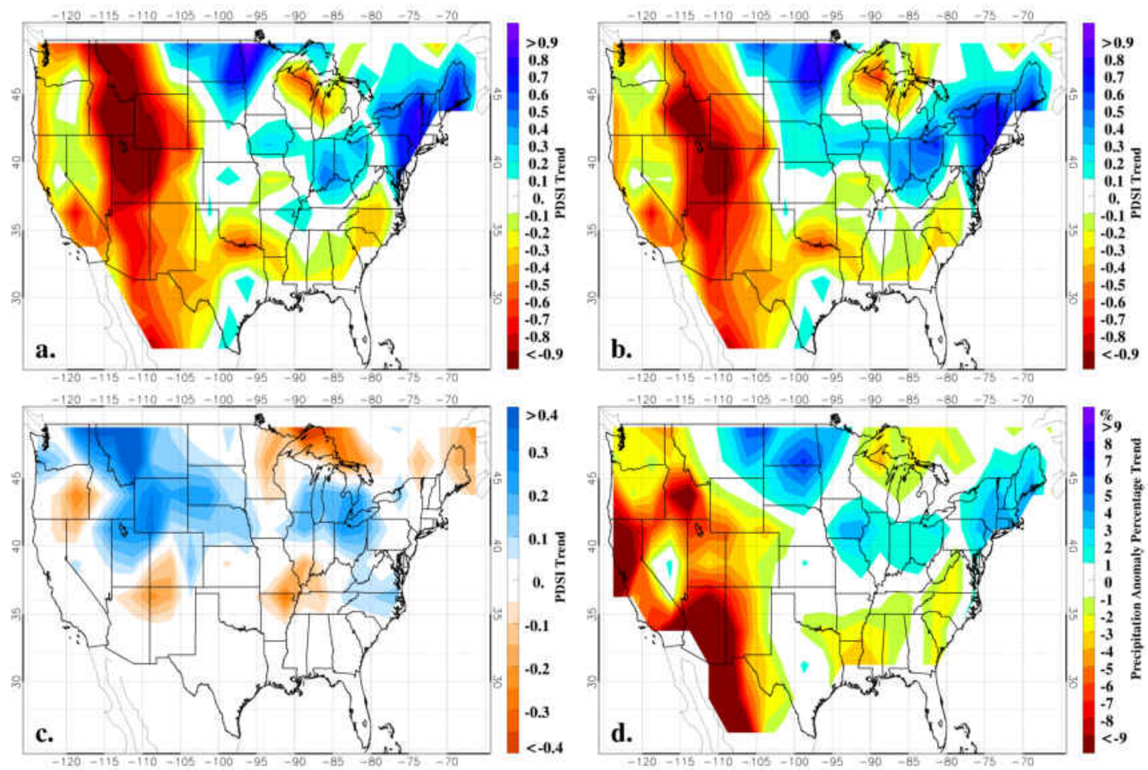

Fig. 22. Trend of PDSI during the period 1979-2010 for (a) the original PDSI model, (b) the modified PDSI model, (c) the difference (modified - original), and (d) the trend of precipitation anomaly percentage relative to the period 1979-2010. PDSI trends are given in PDSI units per decade.

With the modified PDSI redistributing moisture and altering model parameters in winter and spring, the question remains whether this has an impact on the characteristics and trends in the PDSI over the 30-yr period. Fig. 22 presents the trend in PDSI for the two models along with the trend in PAP. Regardless of the model used or PAP, the CONUS is drying significantly in the west, the upper Great Lakes, and to some extent, in the south (Fig. 22a,b,d). Locations with wetting trend include the northeast, southern Great Lakes, and the northern plains. Including snow effect into the PDSI model does little to change the qualitative understanding of these trends (Fig. 22c). The largest impact of this inclusion is a weakening in the drying trend over the northern half of the Rocky Mountains, a moistening trend in the lower Great Lakes, and an increased drying trend in the northern Great Lakes. The inclusion of snow increased the correlation of PDSI to PAP data, from 0.65 to 0.73.

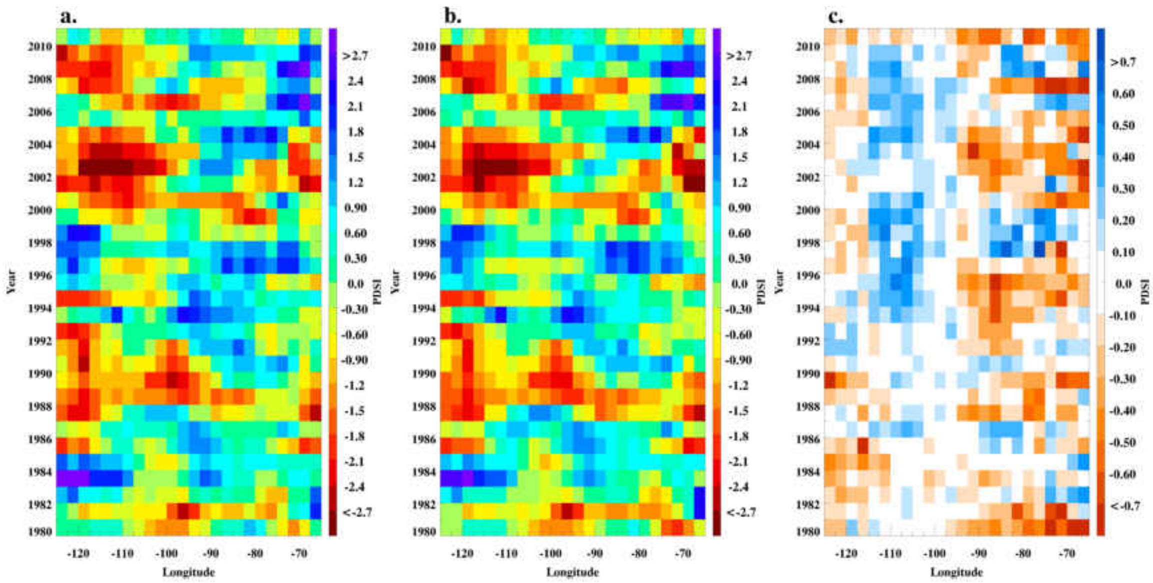

Fig. 23. Annual mean of longitudinally averaged PDSI over the RRB for (a) the original PDSI, (b) the modified PDSI and (c) the difference (modified-original) during the period 1979-2010.

To show the overall moisture conditions over US after adding the snow effect, annual and longitudinal averaged PDSI is presented in Fig. 23. The first year has been omitted due to model spin up (soil moisture is initially assumed to be at capacity; precipitation information is unknown for the previous winter of the first year). Compared Fig. 23a to b, both models show similar moisture condition in CONUS for the past three decades. Well known events such as the drought in 1988, 2002, 2008 and the wet period in 1983 and 1998 due to El Nino are appeared from the outputs of the two models, so does the evolution of each event.

Despite the similarities, the differences between two models are obvious as well. As in Fig. 23c, in general, adding the snow effect made the west of 115° W and the east of 95° W dryer, and the regions in between wetter. Reasons for these differences mainly come from two factors: first, similarly as discussed in model example before, adding snow effect makes the Z index fluctuate more in spring, that is, the Z index is smaller for dry winter and larger for wet winter during the melting period. These fluctuations of Z index make the PDSI decrease for dry periods and increase for wet periods. For example, the relatively high PDSI from the modified model between 115° W and 95° W from 1993 to 2000 is due to the abnormally wet winter in 1993, as shown in Fig. 24. Second, adding snow effect also changes the duration factor, (p and q in Eq. 22), owing to the fluctuations of *Z* index.

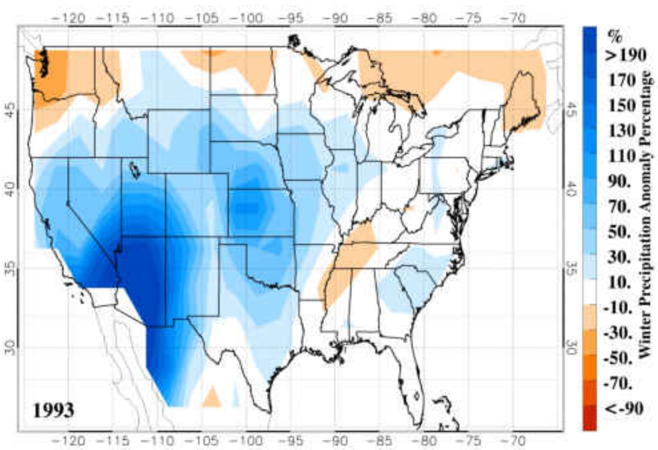

Fig. 24 Winter precipitation anomaly percentage (relative to the corresponding average for the period 1979-2010) for 1993.

Figure 25 shows the difference of q values between the two models for both dry and wet cases. Still using the regions between 115°W and 95°W as an example: q values for dry cases are smaller in these regions (Fig. 25a), therefore, less weight has been put on Z index during dry periods, which makes the current negative Z index value contribute less to PDSI value; similarly, with q values for wet cases larger, more weight has been put on Z index during wet periods, and the positive Z index contributes more to the PDSI value. Therefore, PDSI values will be larger in these regions. Similarly, in the east of 95° W region, q values for dry cases are larger and q values for wet cases are smaller, which makes PDSI smaller in this region.

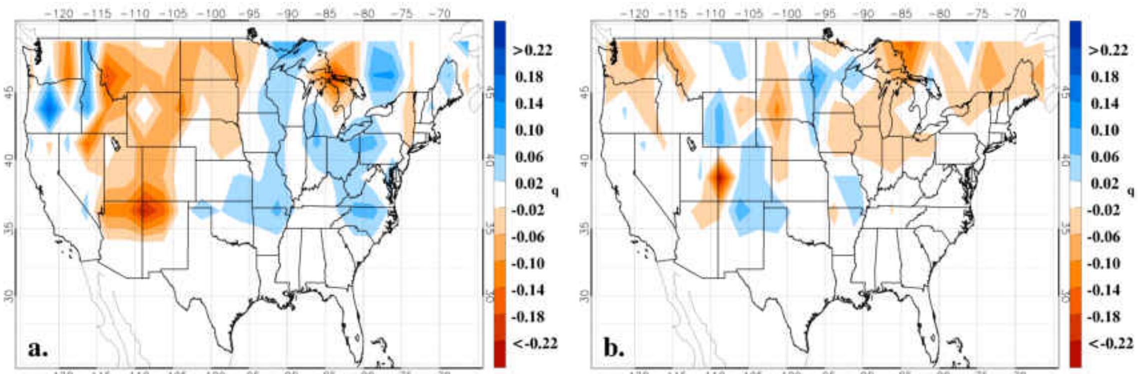

Fig. 25 Changes in the duration factor q (modified-original) for (a) dry cases and (b) wet cases.

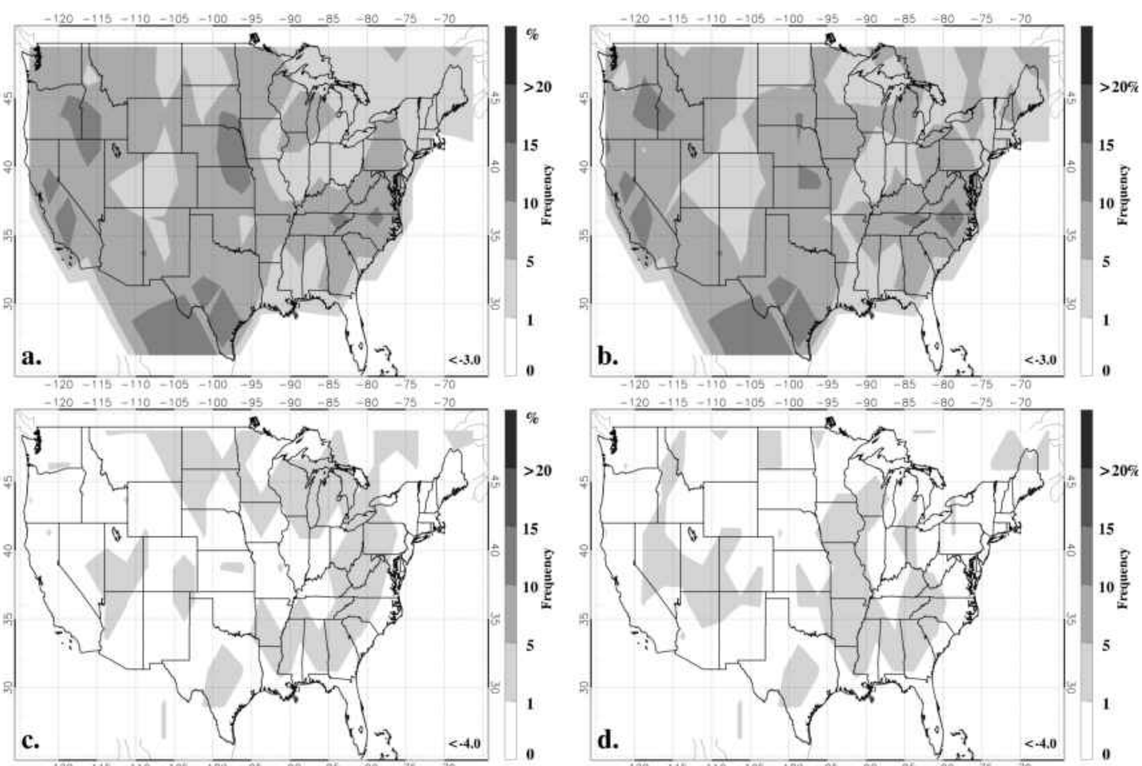

Fig. 26 Probability of PDSI value less than (a) (b) -3.0 and (c) (d) -4.0 for the original PDSI (a) (c), and the modified PDSI (b) (d)

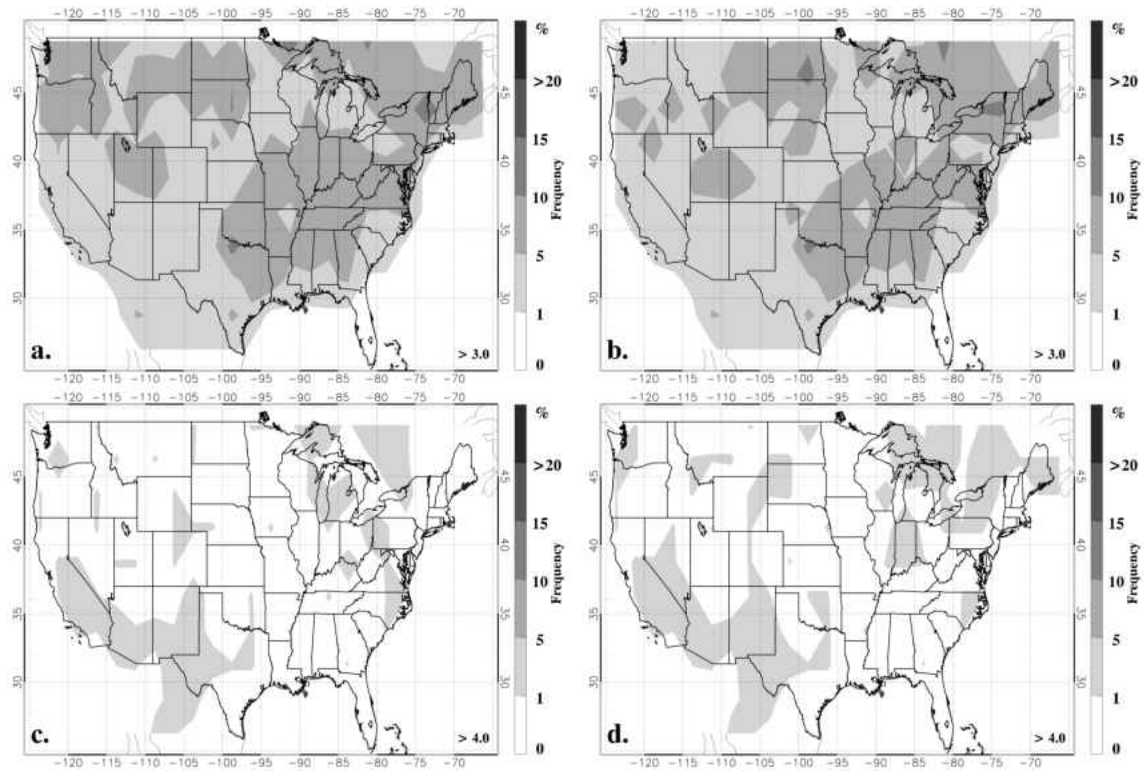

Fig. 27 Same as in Fig. 26, but for PDSI values of bigger than 3.0 and 4.0

With the notion that snow and frozen soil effect make the PDSI more severe for extreme events (both drought and flood), another important properties to check is the probability distribution of PDSI values as studied by Guttman, et al. in 1992. As shown in Fig. 26 and Fig. 27, both the original PDSI and the modified PDSI have 5% to 15% probability for PDSI values < -3 or >3, and the probability for PDSI values < -4 or >4 are mostly below 1%, with some regions range from 1~5 %. The snow effect doesn't have big influence on probability distribution of PDSI values.

Frozen Ground Effect

A lot of efforts have been made to study the influence factors and to parameterize infiltration process over frozen or partially frozen ground. This process, however, is complicated and requires knowledge of a number of variables including: frozen soil depth, snow water equivalent, snowmelt speed and soil properties such as soil temperature, soil type, soil porosity, etc. Measuring these variables is difficult, and nearly impossible using traditional datasets such as reanalyses. Therefore, in this study, frozen ground effect is only been tested for two extreme cases: setting the infiltration rate to 100% over frozen ground, which is the previous setup for the modified model; or setting the infiltration rate to 0% over frozen ground so that all moisture leave the system as surface runoff when ground is frozen. This will provide the opposite extremes for the model and provides bounds for how much PDSI can vary due to the frozen ground effect.

To determine ground status, the MEaSUREs (Making Earth System Data Records for Use in Research Environments) global record of daily landscape freeze/thaw status, version 2 data from NSIDC is used (Kim, Y., et al. 2011). This product is a global record of the daily freeze/thaw (FT) status of the landscape derived from satellite observations of radiometric brightness temperatures it has a spatial resolution of 25 km for almost the entire globe from 01 January 1979 to 31 December 2010.

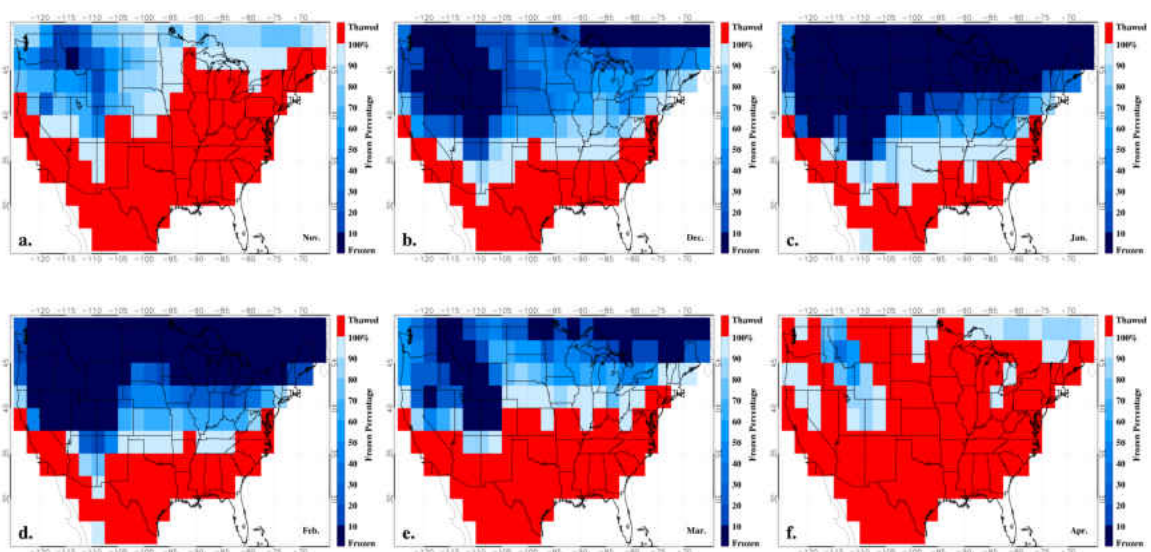

Fig. 28 Percentage of years with ground is unfrozen based on the MEaSUREs (Making Earth System Data Records for Use in Research Environments) global record of daily landscape freeze/thaw status, V2 data from (a) November to (f) April.

Figure 28 shows the climatology of ground status. Start from November, along with the climate mean temperature drop below zero (Fig. 29a), ground start to freeze from the northern mountain area, with ~50% of years ground being frozen; till January, most of the area north of 42° N is frozen. In spring, the ground status lags the air temperature about one month. In March, when climate mean temperature is above zero south of 45° N, the ground status data still have most of the northern area frozen more than 50% of the time. Besides the gap between the two datasets, the climate mean temperature in west coast area is always above 0° C in winter, while the ground status data show the area between 122.5°W and 120°W always frozen in January and February; even the west of 122.5°W area had ~60% of the years frozen. This inconsistency between the two datasets could be caused by the difference in spatial resolution.

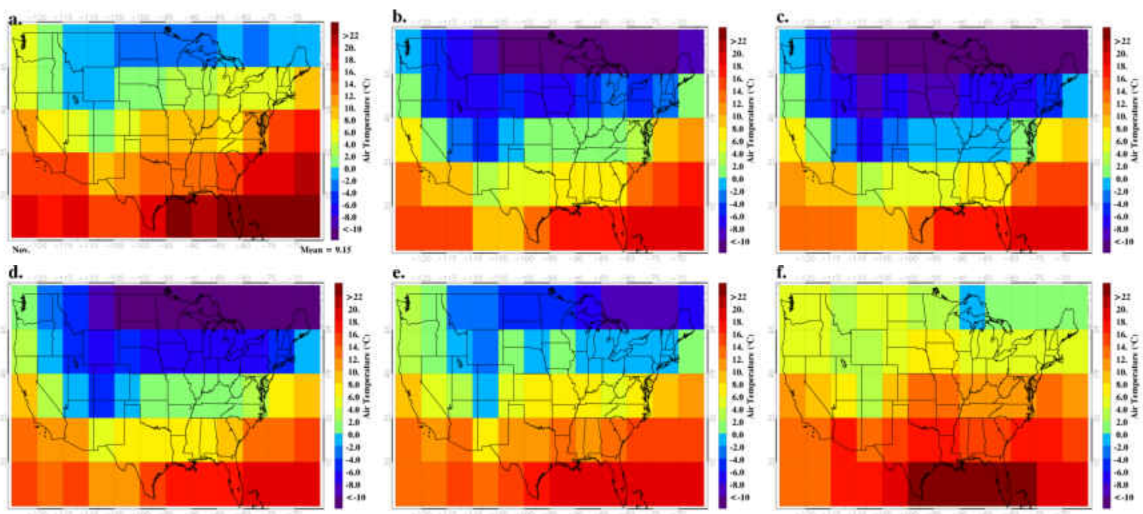

Fig. 29 Climate mean (1979—2010) air temperature from (a) November to (f) April

Due to the lag and the inconsistency between ground status and air temperature data, all the model variables changes a lot in winter and spring. In winter, as ground is frozen in west coast where air temperature is above zero, precipitation didn't enter the system, so that soil recharge decreased while surface runoff increased (Fig. 30b,c). Due to this moisture loss, the climate mean soil moisture decreased >19 cm in the west coast, from ~30 cm in the snow only model to < 10 cm in the model with frozen ground (Fig. 31).

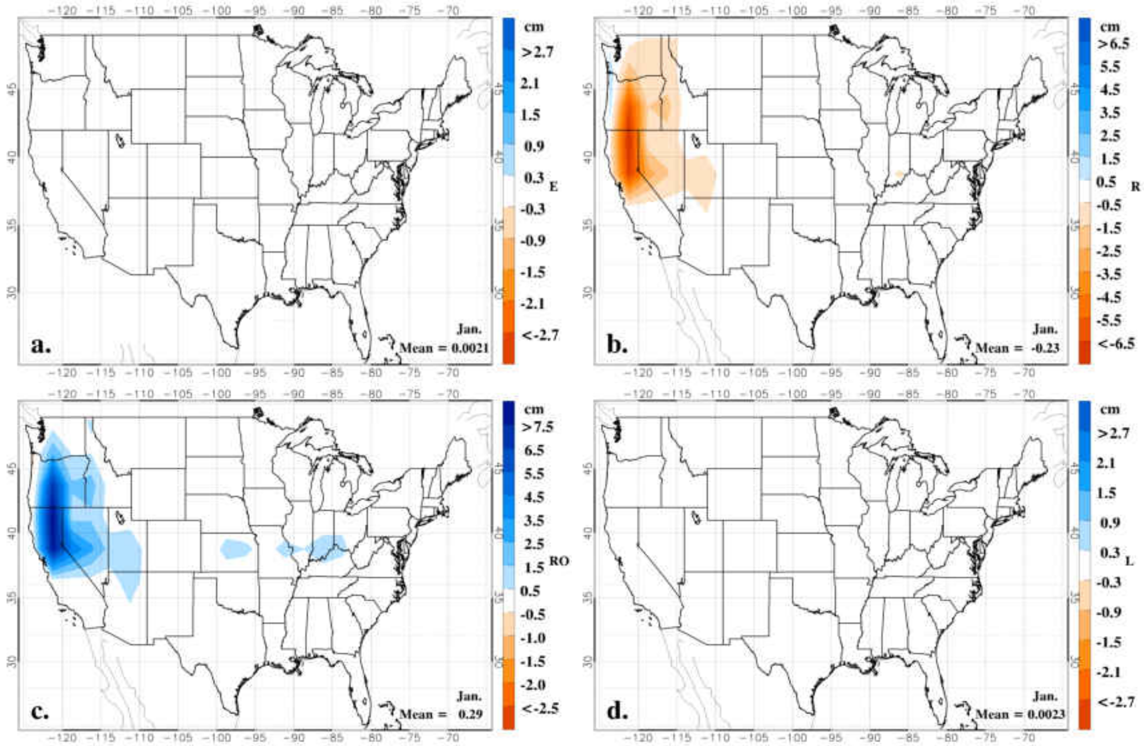

Fig. 30 Climate mean (1979—2010) of the difference between the modified model with frozen ground effect and the model with snow effect only for (a) evapotranspiration, (b) soil moisture recharge, (c) surface runoff and (d) soil moisture loss for January

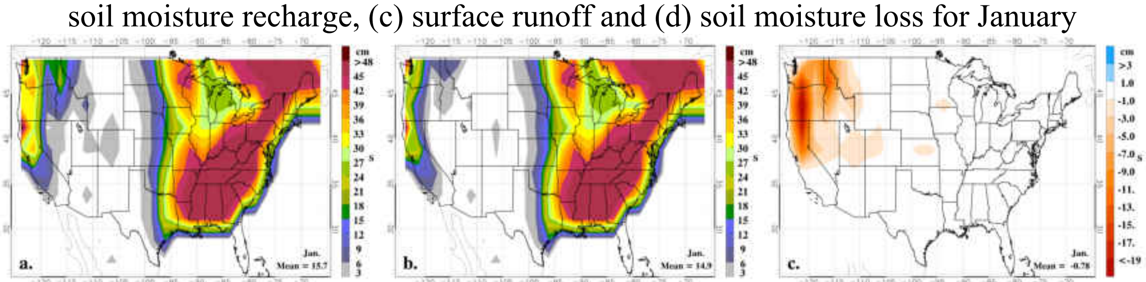

Fig. 31 Climate mean (1979—2010) soil moisture for (a) the modified PDSI model, (b) the modified PDSI model with frozen ground effect and (c) their difference (b-a).

In spring, the largest inconsistency for the two datasets happened in March, snow melt in northern mountain region and the region between 40° N and 45° N (Fig. 32) while ground is still frozen in these regions. Therefore, compared to the snow only model, recharge decreased while runoff increased > 7.5 cm in the northern mountain area (Fig.

33b,c). As the ground status and air temperature is synchronous in northern High Plains and east of United States, all the variables didn't change much in these regions

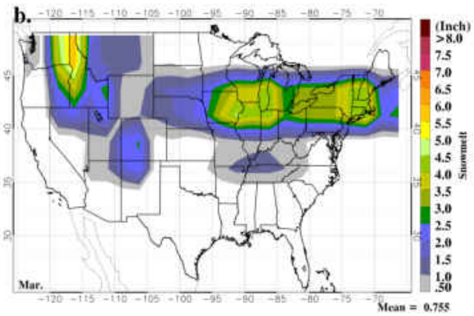

Fig. 32 Calculated climate mean snow melt in March

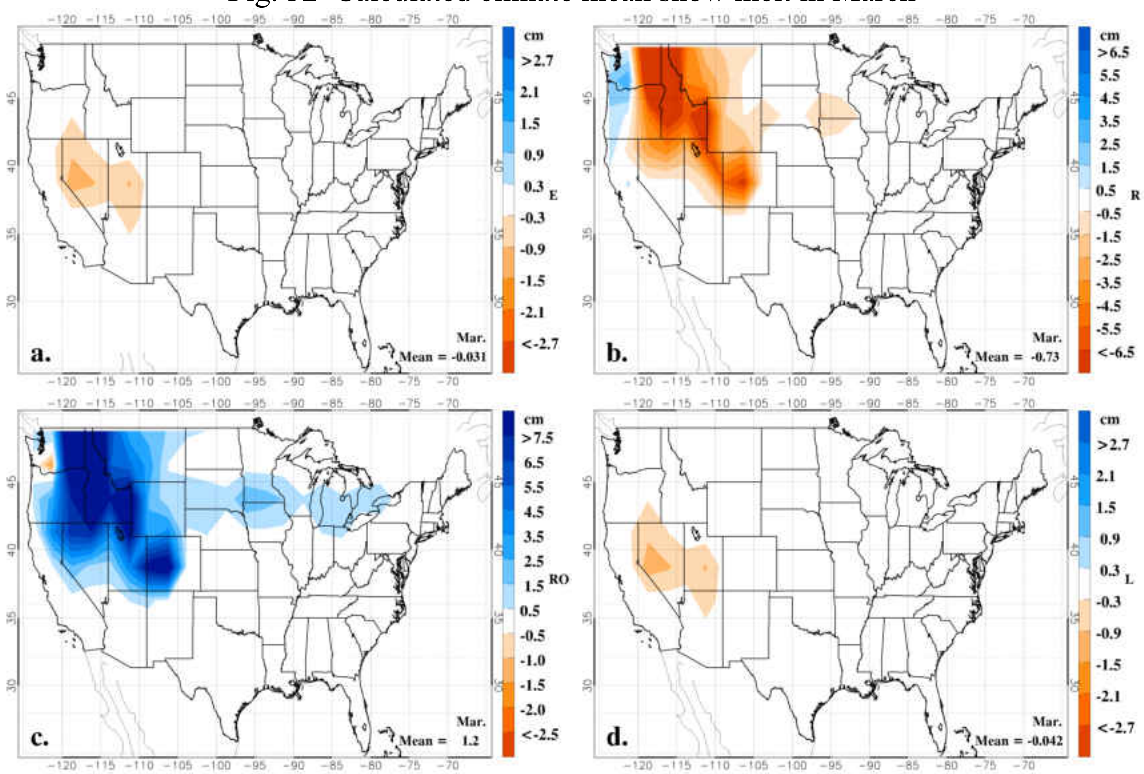

Fig. 33 Same as in Fig. 30, but for March

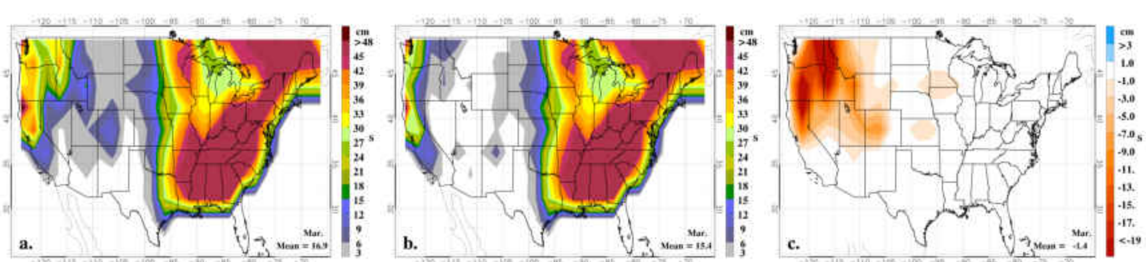

Fig. 34 Same as in Fig. 31, but for March

As precipitation mostly falls in winter in the west coast, and also due to the low evapotranspiration rate in winter, winter and spring are the major seasons for soil moisture to recharge in the west coast region. With frozen ground effect included, the frozen ground block all the winter moisture from entering the system, the climate mean soil moisture in March decreased from 30-35 cm in the snow only model to only <6 cm in the model with frozen ground effect added (Fig 34). This loss in moisture even further influences the model variables in summer. As shown in Fig. 36, the climate mean soil moisture in June in the model with frozen ground is still more than 19 cm lower than the snow only model in the northern Rocky Mountain region; as the rate of evapotranspiration (E) and soil moisture loss (L) all relate to soil moisture, they all decreased in June (Fig. 35).

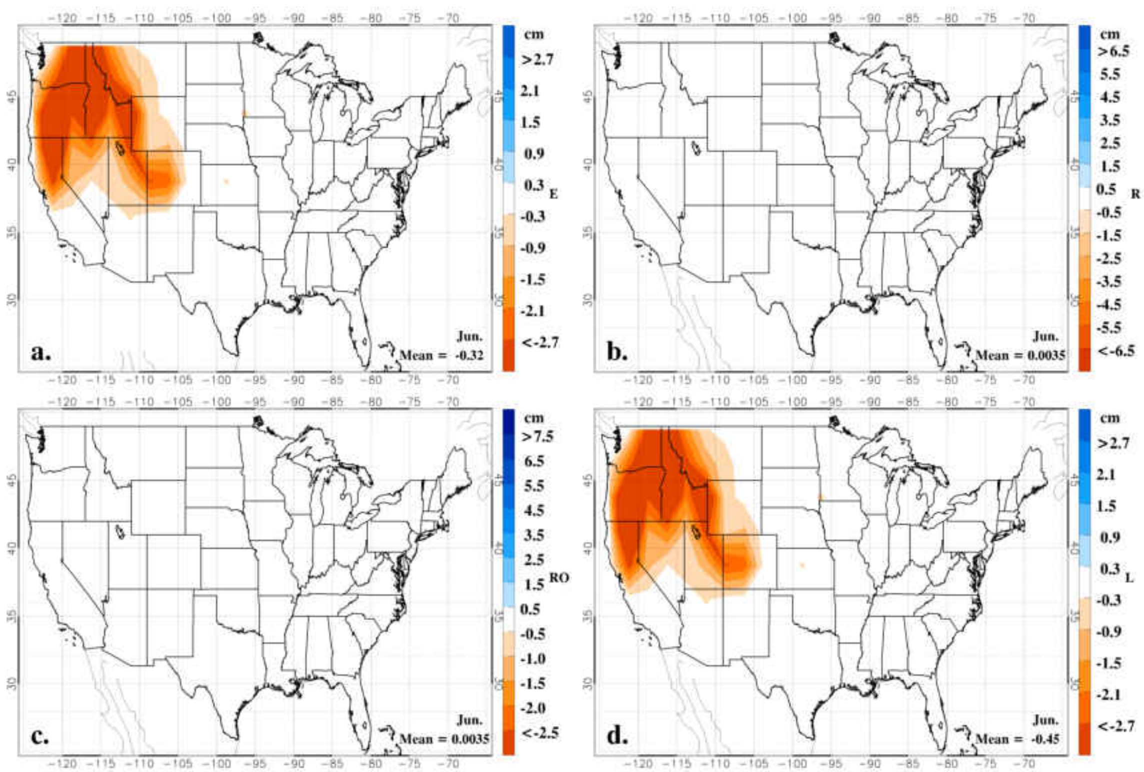

Fig. 35 Same as in Fig. 30, but for June

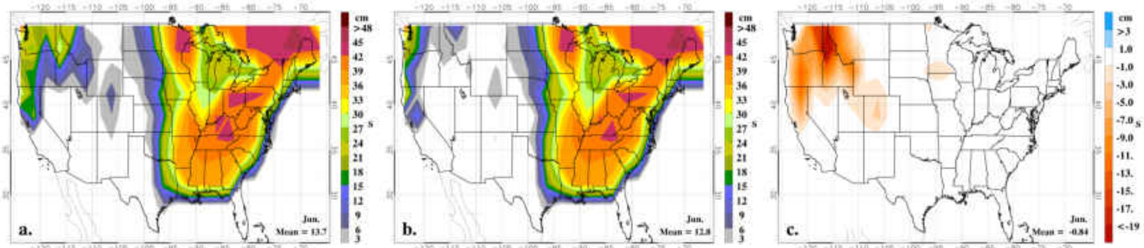
Fig. 36 Same as in Fig. 31, but for June

Therefore, in the snow only model, the moisture only got redistributed over the seasons; soil moisture decrease in winter and increased to the same level in spring. However, in the frozen ground model, the climate mean soil moisture in the Rocky Mountain area decreased from ~30 cm to < 6 cm, which is not realistic. Therefore, based on this research, set the infiltration to 100% over frozen ground is more realistic than completely shut down infiltration over frozen ground. Similar conclusion has been made from the Project for Intercomparison of Land-Surface Parameterization Schemes phase 2(d) experiment (Luo et al. 2002). This project found that over a large region, such as in the grid box of GCM, Hortonian surface runoff can always find its way to infiltrate somewhere in the domain owing to the fact that soil does not freeze homogeneously. Therefore, it was suggested that the inclusion of a frozen soil scheme has little effect on the simulation of soil moisture in large-scale models.

The changes of all the variables in the model with frozen ground effect included can be explained by the following factors. First, low spatial resolution of air temperature data cannot display their variations by the topography. Second, the monthly averaged air temperature cannot show the exact natural lag between air temperature and ground status, which lag varies from a week to a month.

With all the changes in the model variables, the question remained is how much PDSI varied with frozen ground effect. To show the changes, similar plot as in Fig. 23c is made with the difference of the two models. Due to the decrease of soil moisture in Rocky Mountains and west coast region, PDSI value also decreased with frozen ground included; especially for wet years like 1983, 1998, the annual mean PDSI decreased more than 0.7 index value (Fig. 37). For regions east of 105° W, changes in PDSI is minimal.

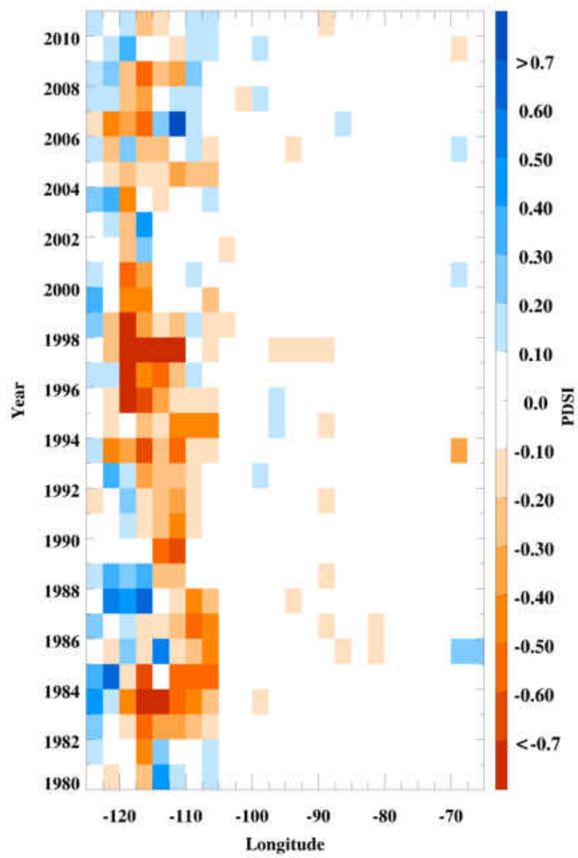

Fig. 37. Same as in Fig. 23c, except for the difference between the modified model with ground status included and the snow only model.

CHAPTER V

CONCLUSIONS AND DISCUSSIONS

Snow processes play crucial roles in both the hydrological cycle and for drought/flood events, and the lack of these processes in the Palmer Drought Severity Index modes has been a major deficiency of this index. This study represents a first attempt to include snow and ground status into the Palmer Drought Severity Index model. Following the original model setup in the PDSI, precipitation has been withheld and accumulated as snowpack on the ground based on the air temperature threshold, and all the other processes have been shut down in winter. Modifications also have been made in the calculation of CAFEC precipitation and moisture departure as well. CAFEC precipitation has been replaced by the climate mean precipitation in winter so that moisture departure in winter in the modified model simply is precipitation anomaly. In spring, snowmelt calculation is based on the degree-day method, and the snowmelt for the whole season has been redistributed into one month and released into the system as moisture input, so that the moisture departure is calculated as the precipitation plus snowmelt minus the CAFEC precipitation in spring melting season.

With snow effects included, all model variable, soil recharge and loss, surface runoff are zero in winter and then the spring melt recharge the soil in west of United States while most of them became runoff in the east of US. The snowmelt also made the CAFEC precipitation increased more than 200% in the melting season in northern latitudes and mountainous areas. The correlation between the increase in CAFEC precipitation and snowmelt are 0.98 or higher in melt season.

In the aspect of monitoring extreme events with the index, including snow effect better represented spring flooding events at northern latitudes that were caused by snowmelt. For example, for the 1997 spring flood in Grand Forks, the Z index changed from ~1 in the original model to ~5 index value in the modified model in April 1997. Despite the large value of Z index, PDSI value only change ~1 for that month but this increase in PDSI lasted almost a year, which further prove that PDSI is designed for long-term impact, such as drought, and is not suitable for short-term and intense events such as flood. When applying to drought events, the modified PDSI changed the duration factor in Colorado region and better represent the actual drought condition.

On a climatic basis, the inclusion of snow effect causes minor changes to the climate mean PDSI values over the CONUS. However, the snow effect made areas west of 115°W and east of 95°W dryer while the regions in between became wetter during the past three decades. These changes are mainly caused by the changes in the weighting factor. Both the original and modified models capture an overall drying trend in the western US, especially in the mountainous area and the desert southwest. The northeast, southern Great Lakes, and the northern plains regions of the US are getting wetter during the past three decades. Further more, the inclusion of snow had no major influence on the spatial comparability of PDSI since both models showed similar probability distribution as in Wells et al. (2004).

Finally, the effect of frozen ground on the PDSI model and on the index has also been tested. A compared run was made with setting the infiltration over frozen soil to 0% and 100%. Due to the inconsistency as well as the time lag between air temperature data and ground status data, the frozen ground effect caused most of the snowmelt in west coast and northern Rocky Mountain region lost as runoff. And because of this moisture loss, soil moisture the in model with frozen soil reach unrealistic low level and further caused evapotranspiration and soil moisture loss decreased in these regions for all seasons. Therefore, this research suggests that when applying to the large scale and monthly time step PDSI model, setting the infiltration to 100% over frozen ground is more realistic than completely shut down infiltration over frozen ground.

Future work

Besides the improvement in PDSI, this study also raised a number of issues that warrant further investigations. The complexity of the snowmelt process requires adaptation of the PDSI model to a weekly interval and higher spatial resolutions to better capture this phenomenon. The partitioning of snowmelt between runoff and infiltration over frozen ground is a major knowledge gap, and in the future, this process should be parameterized in detail to allow for fractional amounts of infiltration. Finally, the weighting factor originally derived in Palmer (1965) should be modified to the self-calibrated method (Wells et al., 2004) to better represent the temporal and spatial characteristics of present-day studies.

APPENDIX

Appendix A

List of Symbols and Acronyms

a	Degree-day ratio (mm °C$^{-1}d^{-1}$)
α_{mx}	Maximum degree-day ratio assumed to occur on June 21 (6.0 mm °C$^{-1}d^{-1}$)
α_{mn}	Minimum degree-day ratio assumed to occur on December 12 (2.0 mm °C$^{-1}d^{-1}$)
c_p	Specific heat capacity of air ($c_p = 1005\, J\, kg^{-1}\, K^{-1}$)
d	Moisture deficiency/excess (*mm*)
e_s	Saturated vapor pressure of air (Pa)
e_a	Vapor pressure of free flowing air (Pa)
δe	Vapor pressure deficit, or specific humidity (Pa)
g_a	Conductivity of air, atmospheric conductance ($m\ s^{-1}$)
p	Duration factor
q	Duration factor
$[AWC]$	Available water capacity of both soil layer (*mm*)
CAFEC	Climatically Appropriate For Existing Condition
E	Evapotranspiration (*mm*)
K	Weighting factor

K'	Weighting factor
L	Soil moisture loss (*mm*)
M	Snowmelt rate (*mm/d*)
P	Precipitation (*mm*)
$\hat{P}$	CAFEC precipitation (*mm*)
PAP	Precipitation anomaly percentages (%)
$[PE]$	Potential evapotranspiration (*mm*)
$[PL]$	Potential Loss (*mm*)
$[PR]$	Potential recharge (*mm*)
$[PRO]$	Potential surface runoff (*mm*)
$[PL_s]$	Potential loss from the surface layer (*mm*)
$[PL_u]$	Potential loss from the underlying layer (*mm*)
R	Recharge to soil (*mm*)
R_n	Net irradiance ($W\ m^{-2}$)
$[RH]$	Relative humidity
$[RO]$	Surface runoff (*mm*)
RRB	Red River Basin
S	Soil moisture for both layers (*mm*)
S_s	Available moisture stored in surface layer (*mm*)
S_u	Available moisture stored in underlying layer (*mm*)
T	Degree days, (℃ d)
T_a	Air temperature (℃)
T_m	Melt threshold temperature (0 ℃)

T_{SN}	Snowpack temperature (°C)
Z	Z index
α	$= \overline{\mathrm{E}}/\overline{[\mathrm{PE}]}$, coefficient for evapotranspiration
β	$= \overline{\mathrm{R}}/\overline{[\mathrm{PR}]}$, coefficient for soil moisture recharge
γ	$= \overline{[\mathrm{RO}]}/\overline{[\mathrm{PRO}]}$, coefficient for surface runoff
ε	$= \overline{L}/\overline{[PL]}$, coefficient for soil moisture loss
θ	Psychrometric constant ($\theta \approx 66 Pa\ K^{-1}$)
λ_v	Latent heat of vaporization. Energy required per unit mass of water vaporized. ($J\ kg^{-1}$)
ρ_a	Dry air density ($\rho_a = 1.225\ kg\ m^{-3}$)
ρ_S	Density of snow ($kg\ m^{-3}$)
ρ_W	Density of water ($kg\ m^{-3}$)
σ	Snowpack temperature lag factor (0.5)
Δ	Rate of change of saturation specific humidity with air temperature ($Pa\ K^{-1}$)

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