Hydrological indices and triggers, and their application to hydrometeorological monitoring and water management in Texas

George H. Ward
Center for Research in Water Resources
The University of Texas at Austin

Final Report

TWDB - UTA Interagency Contract No. 0904830964

Project Officer: Mark Wentzel, Ph.D.
Surface Water Resources Division
Texas Water Development Board

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Executive Summary

Assessing and responding to variations in hydroclimate require some immediate quantitative measure of surface-water conditions, not only of surfeits or deficits in rainfall, of volumes of flow in the streams and rivers, and of the storage in reservoirs, but how these relate to normal conditions and their likelihood of meeting regional water demands. A metric that summarizes surface-water conditions in a compact and intuitive way is an indicator or index (the latter generally being more mathematically complex). Specific values, or thresholds, of an index can be associated with levels of surface-water availability and, in turn, tied to specific management actions, in which case they are referred to as "triggers."

The overall objectives of this project are to survey the literature on hydroclimatological indices, with emphasis on the international literature for case studies of trigger formulations, then to use Texas examples to explore and contrast different indices of hydroclimate and their thresholds. The indices specifically addressed in this review are selected because either they appeared to offer some potential for utility in Texas water management, or because they have acquired some level of acceptance in hydrometeorology and water-resources.

An *indicator* is a measured parameter, or a statistic of a set of measurements of the parameter, that serves to track the variation of the state of some complex entity. In the present report, an *indicator* is a directly measureable variable or an associated statistic, while an *index* is a numerical aggregation of one or several indicators. For purposes of identifying candidate indices for potential utility in Texas, three general criteria were applied. First, values of the index should be routinely and readily available on at least a monthly basis, or easily computable from variables that are available. Second, there should be extant a long period of record of the index to serve as a foundation for hydroclimatological analyses. Third, there should be good evidence in the literature of the actual or potential applicability of the index to the Texas environment. In addition to these criteria, the qualitative attributes of appropriateness and communicability, along with the technical basis of the index formulation, were considered to further winnow the field of choices.

For analysis of a period of record of an index and for its operational application, it is convenient to have a companion specification of threshold values, truncation levels, or triggers that categorize moisture conditions, for purposes of appraising the statistics of occurrence of such conditions and conveniently displaying those results. In extremes of flood and drought, such trigger values carry an implication of specific mitigative actions, such as water restrictions, dam releases, or other management responses. A clear, intuitive index that is objective and readily disseminated to the public has obvious operational advantages. This study focuses upon indicators and indices of hydrometeorology, mainly the variables of precipitation and evapo-transpiration, of streamflow, and of reservoir contents, with an implicit averaging time of weeks to months. The most convenient sources of routinely updated data are the climatic division statistics of the National Climatic Data Center (NCDC), and daily streamflows from the gauge data of the U.S. Geological Survey (USGS). The three variables, rainfall, streamflow, and reservoir contents, are used to organize and present the results of the literature survey.

Hydrometeorological parameters, when used as the basis for an indicator, are typically expressed as an anomaly time series, that is, a time series of departures from a reference value or condition. Definition of the reference condition becomes part of the formulation of the indicator, depicting some measure of "normalcy." The simplest such reference is a long-term mean, but the reference condition can be more complex. Sometimes, the anomaly is standardized. This is a way of scaling the anomaly: the reference condition is used to define a mean and standard deviation, and the departures from that mean are divided by the standard deviation.

Practical application of an index often focuses on segments of the time series in which the index is above or below its reference condition(s). In this case, it is not the individual value of the index that is of concern so much as its systematic persistence of surfeit or deficit. Various terminologies have been employed historically for these periods in a moisture-index time series, including "wet spell," "dry spell," "pluvial" and "drought." Identification of such periods clearly requires a definition of when they begin and end, which in turn necessitates threshold or trigger values of the index (or indices) for each type of condition, perhaps supplemented by other identification rules.

Table A
Moisture condition decile classes of annual and monthly precipitation used by Australian Bureau of Meteorology, from Kininmonth et al. (2000)

designation	*cumulative frequency value*
very much below average	≤ 10%
below average	10 – 30%
average	30 – 70%
above average	70 – 90%
very much above average	90 – 100%

Because of its simplicity of measurement and availability of data, precipitation has been used historically as a climate indicator. With a suitably defined reference value $\hat{P}$, the precipitation anomaly $P_i - \hat{P}$ (where P_i is the rainfall in month *i*) is a fundamental indicator for hydroclimate. $\hat{P}$ is almost always a long-term mean, either the annual mean or monthly mean, and is often a climatological normal. The reference value itself can be the threshold (i.e., above-normal or below-normal). A slightly more complicated indicator is the precipitation time series with multiple thresholds. An example of an index based upon precipitation is the use of cumulative frequencies of occurrence, called the precipitation-decile index. The time series, usually monthly, is re-ordered by magnitude, from which the cumulative frequency becomes the index, and the decile points are identified as thresholds. This index has been used in Australia for many years as a basis for monitoring and reporting moisture conditions, see Table A.

The standardized precipitation index (SPI) has become popular since its introduction in 1993 as a metric of drought conditions in Colorado. There are two components to the calculation of the SPI. The first carries out a calculation of the accumulated precipitation over a sliding time window of selected duration, referred to as the "time scale." The SPI with time scale of *M* months is conventionally designated SPI-*M*. The second standardizes the cumulative precipitation values for a given time scale, relative to a long-term mean-cumulative, to units of standard deviation, employing a mathematical transformation to the standard Gaussian distribution (with zero mean and unit standard deviation). In other words, the SPI-*M* takes a

Table B
Moisture condition classes for standardized precipitation index (SPI), from McKee et al. (1993, 1995)

designation	SPI *value*
extremely wet	≥ 2.0
very wet	1.5 – 2.0
moderately wet	1.0 – 1.5
near normal	-1.0 – 1.0
moderately dry	-1.5 – -1.0
severely dry	-2.0 – -1.5
extremely dry	≤ -2.0

record of monthly precipitation, creates a new record of *M*-month sums, and produces an index that generally ranges ± 3 about 0 (though theoretically its range is unlimited). The moisture-class thresholds for the SPI in conventional use with this index are shown in Table B.

The range and behavior of the SPI-*M* are greatly dependent upon the selection of accumulation time *M*. This feature is considered by many to be the primary advantage of the SPI, because it allows the selection of a time scale for investigation to be entirely in the hands of the user. This in effect is equivalent to subjecting the precipitation time series to a moving average of prescribed duration. Longer window durations will have the effects of smoothing the input time series and lagging it in time.

The Palmer index – which is in fact several indices – is a general parameter for tracking moisture availability at the surface of the ground, including both wet and dry conditions. The most familiar form is referred to as the Palmer drought severity index (PDSI). The complexity of the index, its physical basis, and its properties derive in part from the definition of the hydro-climatological reference value $\hat{P}$, referred to as the precipitation *climatically appropriate for existing conditions* (CAFEC). The index as well as the CAFEC is based on a rudimentary soil-water budget, and includes soil-water deficits and surfeits driven by evaporation and transpiration, as well as rainfall deficits and surfeits. Not only is its computation based on a time series of a complicated soil-water accounting, but the reference value $\hat{P}$ for the moisture

Table C
Moisture categories for Palmer drought severity index (PDSI), from Palmer (1965)

designation	*PDSI value*	*designation*	*PDSI value*
extremely wet	≥ 4.0	near normal	-0.5 – 0.5
very wet	3.0 – 4.0	incipient drought	-1.0 – -0.5
moderately wet	2.0 – 3.0	mild drought	-2.0 – -1.0
slightly wet	1.0 – 2.0	moderate drought	-3.0 – -2.0
incipient wet spell	0.5 – 1.0	severe drought	-4.0 – -3.0
		extreme drought	≤ -4.0

anomaly is itself a time series, with monthly values dependent upon "potential" values of the terms in the soil water budget. The precipitation anomaly $P_i - \hat{P}$ is multiplied by an empirical scaling (or "weighting") factor whose purpose is to "normalize" the moisture anomaly for the climatological region, to arrive at the *Z*-index. The Z_i values are accumulated in time, and the resulting cumulative anomaly scaled to ± 4 based on extreme droughts in the Midwest during the Dust Bowl years. These moisture categories are presented in Table C. This results in the X-index (in Palmer's original designation). Overlaid on the index formulation are protocols of determining a "dry spell" or "wet spell" in which the X-index is re-initialized at the beginning of such a spell. This index together with these protocols is the PDSI.

The Palmer index, particularly the PDSI, was a monumental step forward in the analysis and monitoring of drought in the United States and elsewhere. It has been extensively reviewed in the literature, and several deficiencies have been identified, one of the most serious to be that the PDSI was developed for a limited part of the Midwest, with a modest extension to encompass nine climatic divisions in seven states. There is little evidence that the weighting factor, designed to normalize the index for other climates, is adequate to the task. Indeed, Palmer's original intent to scale the index to the range -4 to +4 clearly fails for other geographic regions of the U.S., including Texas, where its range is nominally ± 6 and greater. Globally, it ranges ± 10.

Palmer's method for identifying the start and end of a wet spell or dry spell has been described by several critics to be "rather arbitrary," one result of which is that the PDSI exhibits unrealistic variations from one month to the next. This is apparent in the examples from Texas. The procedure of re-initializing the X-index at the beginning of a dry or wet spell has two effects. The first is to completely alter the utility of the categories (Table C) as thresholds of moisture conditions. The second, and perhaps more important, is to disrupt the progression of the response of the index to the changing soil water budget. This was indeed the objective of Palmer, finding that the X-index did not seem to respond as quickly as he wished to changing moisture conditions. Of the indices reviewed in this study, this procedure is unique. The usual strategy is to define the index however appropriate, then use thresholds to categorize the resultant moisture conditions, including anomalous periods. No other index is modified in the course of determining the beginning or end of a wet or dry spell. There are numerous such restarts in the PDSI time series. In the Texas NCDC climatic divisions, these average more than once a year over the 1895-2011 period.

The notion that aridity can be measured by the degree to which precipitation P fails to satisfy the demand for water, as measured by potential evapotranspiration PE, dates back nearly a century, and has resulted in several indices that include both precipitation and potential evaporation. Thornthwaite proposed a moisture index given by $(P/PE - 1)$. Similar indices were used in the mapping of arid regions by the United Nations Educational, Scientific and Cultural Organization (UNESCO), and by the United Nations Environmental Programme (UNEP). The UNEP effort defined the aridity index to be P/PE. This index was used in a global mapping of drylands, and has been employed to track increased desertification across the earth.

One promising new proposal is the standardized precipitation-evapotranspiration index (SPEI), which is essentially identical to the SPI except including PE in the anomaly definition. It was not selected for evaluation for Texas because it is not routinely calculated for the state and is too complex to be undertaken by Texas water managers. However, a similar index was proposed in 2010 for use in the Colorado basin (of the West), called the hydroclimatic index (HI). Like the SPEI, this index computes a monthly time series of the difference $(P_i - PE_i)$, and aggregates the monthly data into windows of 3-, 6-, 12-, 24-months, etc., exactly like the SPI, but the underlying mathematics are simpler and the index is capable of easy implementation in spreadsheet software.

In many respects, streamflow would appear to be a more appropriate basic metric for water supply than precipitation. There are fewer such indices in the international literature, however, possibly because the data record is more limited for streamflow than precipitation. Most of these indices are simple flows above or below a reference value. As is the case for precipitation, it is natural to define a streamflow index as a standardized anomaly, e.g., the ratio of departure from the mean (or some other suitable reference) to the standard deviation, and there are several examples in the literature. The success of the SPI has motivated an analogous index for streamflow, in which monthly flows are accumulated in a window of specified length, fitted by an appropriate cumulative distribution function, and transformed to a standardized Gaussian. Several versions of this have been proposed, such as the standardized runoff index (SRI), the standardized flow index (SFI), the standardized streamflow index (SSI), and the streamflow drought index (SDI). None of these is appropriate for use in Texas because of the complexity of the calculation and the limited experience with the index.

A streamflow-based method particularly useful for diagnosis and display of the occurrence of pluvial and drought periods is provided by the so-called residual mass curve (RMC), given by the cumulative sum $\Sigma\,(Q - \overline{Q})$ where $\overline{Q}$ is the period-of-record mean. A period of below-average flow is exhibited in the time plot of the RMC as a declining trend in the curve. Similarly, a period of above-average flow is a rising trend in the curve. The trend can be then be quantified as the least-squares line passing through the RMC data for each period identified. The main utility of this approach is that it provides a convenient, intuitive procedure for detecting anomalous-moisture periods. The duration is defined by the time from the first point of the rising or declining segment to the intersection with the mass curve of a line of pre-defined slope. The steeper the slope of the regression line, the more intense the pluvial or drought in terms of average flow surplus or deficit.

This method has been recently applied in Texas and shows promise. The method was suggested to the Guadalupe-San Antonio Basin and Bay Expert Science Team (BBEST) during its work on formulating flow standards for the basin (which overlapped with the period of study of the present project). Interest was, of course, focused on drought events. In this case, a drought was

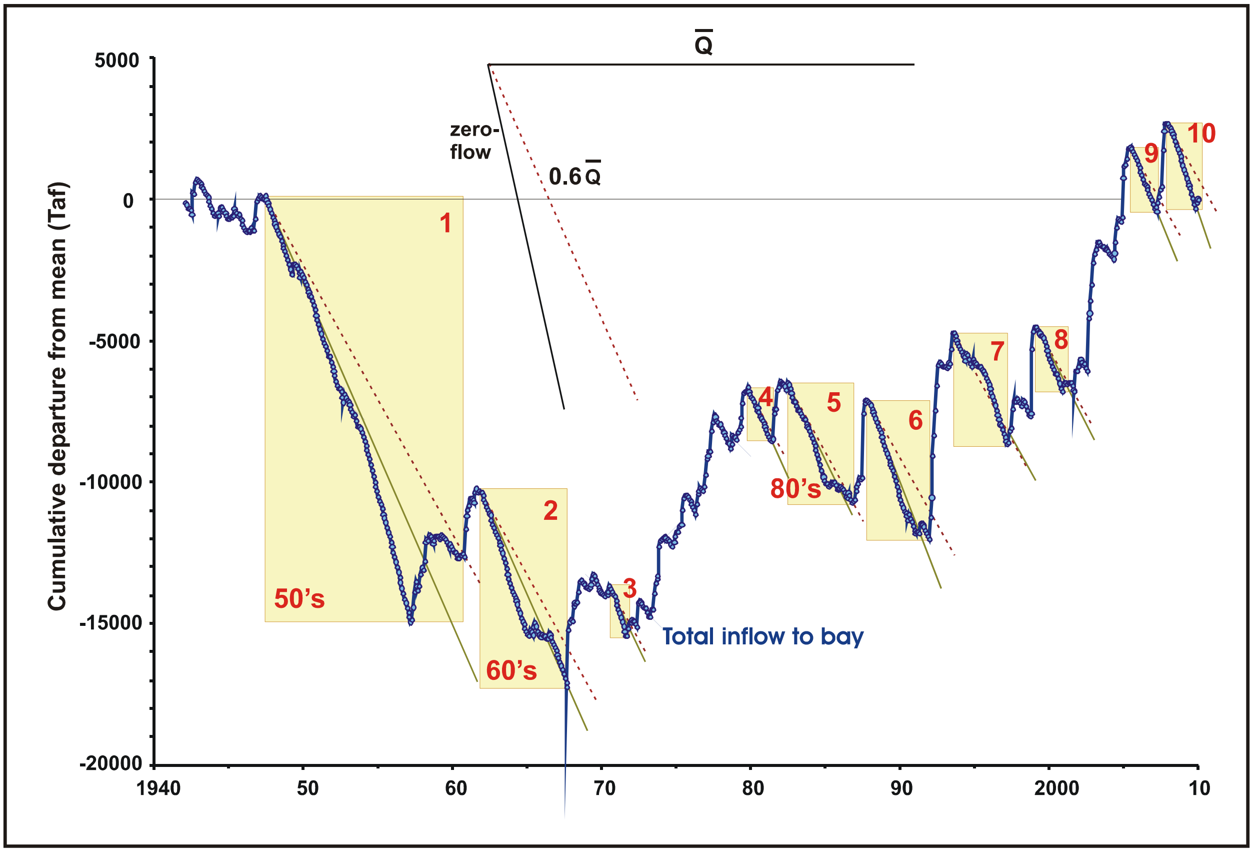

Figure A - Residual mass curve of monthly inflow time series of San Antonio Bay, prominent droughts indicated by constant-flow (0.6 $\overline{Q}$) criterion lines and regression lines

defined to be referenced to 60% of the period-of-record mean flow. The RMC time plot is shown in Figure A, based on the 1942-2009 record of monthly flows into San Antonio Bay (the combination of gauged flows in the San Antonio and Guadalupe Rivers and estimates of nongauged runoff). Ten drought periods are detected, numbered from 1 to 10 in Fig. A, of which 1 is the notorious Drought of the 50's, and 2 is the drought of the 60's. TWDB has authored an EXCEL® application with VBA Macro to compute this index. In addition to the above application, the TWDB EXCEL implementation has been recently applied to evaluating the effects of drought on the Texas estuaries (Montagna and Palmer, 2012). However, further research is needed to develop this methodology for general use, because it is presently not usefully scaled and the method of drought/pluvial determination needs refinement.

Table D
Hydroclimate indices selected for potential application to Texas watercourses

(a) short-term memory	SPI-6	HI-6	
(b) one-year memory	SPI-12	HI-12	surfeit N index
(c) long-term memory	SPI-24	PDSI	Palmer X index
	RMC		

In some respects, the contents of water in a reservoir can be viewed as the most basic metric for water management. For a supply reservoir, the contents relative to probable inflow measures the vulnerability of the water supply, and in a drought condition the contents represents the immediate availability of water to meet demands. There are numerous cases of extensive reservoir development in the international literature, some of which are in basins with hydroclimatologies similar to those in Texas, but in every case, water management during drought devolves to locally implemented water restrictions, and is not linked to thresholds of an index. The literature review for examples in which basin-wide water management is tied to reservoir levels, or cases in which reservoir management is tied to triggers of a hydrometeorological index, proved futile.

Nine candidate indices were selected for further evaluation for applicability in Texas, listed in Table D, though the two short-term indices are expected to have only limited usefulness. The PDSI and SPI were obvious choices from their wide acceptance and historical application. Further, monthly updates of these indices for the state climatic divisions are routinely available from the NCDC. Given that it is usually the longer-term vacillations in hydroclimate of concern in Texas, the SPI-12 and SPI-24 were deemed most suitable as candidate indicators. The SPI-6 was also chosen because occasionally there may be use for an index with short-term responsiveness. The often erratic behavior of the PDSI due to the re-initialization employed as part of the wet/dry-period protocol is undesirable, however. As an alternative, the Palmer X-index may have merit, as it is basically the PDSI without these protocols. This is readily calculated using the Z-index, which is also a routine product of NCDC. The residual mass-curve

(RMC) index using monthly streamflow has already found some utility in Texas applications, and is readily calculated from streamflow data (or by using the TWDB EXCEL product).

The surfeit of annual precipitation as a fraction of mean precipitation was selected as a simple index with historical use. Specifically, the index is $N_i = (\overline{P}_i - \hat{P}) / \hat{P}$ where $\overline{P}_i = \sum_{k=i-11}^{i} P_k$, and $\hat{P}$ is the normal annual precipitation, for this application taken to be the 1971-2000 normal. Here N_i is a surfeit fraction-of-normal of the 12-month mean ending in the current month i, so it varies monthly. This is simply calculated from the NCDC division-mean monthly precipitation.

As noted above, the SPEI is not routinely available nor easily calculated, nor has it yet received the validation of widespread use. To explore the potential of this sort of index, which measures the excess of precipitation over surface demand by $P - PE$, the much more facile hydroclimatic index (HI) was selected instead. This is readily calculated from NCDC division-mean rainfall and temperature data, and the Thornthwaite equation for potential evapotranspiration. For comparability to the SPI-12 and the N index, a 12-month window was used. For comparability to the SPI-6, the HI with a 6-month window was also selected. These are designated HI-12 and HI-6, resp., in Table D.

Eight demonstration cases were set up to compare the performance of these nine candidate indices. These were USGS gauge sites whose watersheds were entirely or predominantly in a single NCDC climatic division. Gauges were sought on major rivers, exhibiting a range of watershed climatologies, and minimal impacts from upstream reservoirs (though in Texas some upstream reservoirs cannot be avoided on major rivers). The data source for the RMC analysis is the monthly mean streamflows developed from the USGS daily data for the selected gauge. The NCDC division-mean values of precipitation, temperature, SPI and PDSI (and its Z-index) are used. The period adopted for the test cases is 1945-2011, to include a range of surface-water conditions, notably the Drought of the Fifties, and the intense pluvials and droughts that have occurred in recent years. (For a few of the streamflow gauges, the records only extend back to the late 1950's.) Plots are provided for 20-year intervals, starting in 1950, and are organized by index memory, as grouped in Table D.

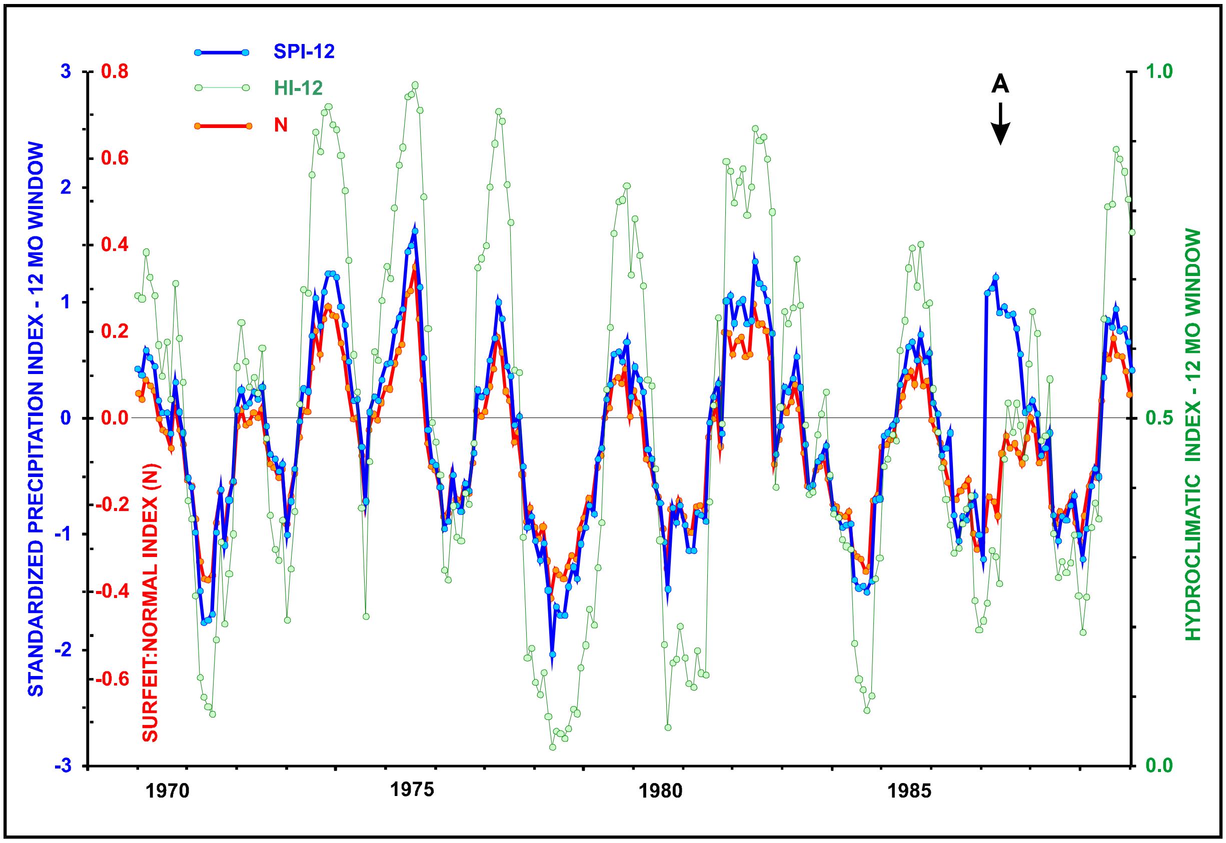

Figure B - Sulphur River at Talco moderate (1 year) memory indices (see Table D), for 1970-1989

It is noteworthy that the highest value of all of the pairwise correlations is between N and SPI-12, ranging 0.97-0.98 over all the gauges and divisions tested. This means that, apart from scaling, the two indices are essentially identical. Despite its apparent sophistication, the SPI-12 does not improve over a simple 12-month average referenced to the mean normal rainfall. This also suggests that both indices are equally capable of normalizing for geographic variation. Despite this high correlation, there is occasional aberrant behavior, particularly of the SPI-12. An example is the response of SPI-12 during the late 80's drought, in which it declares a non-drought, while the other indices continue to indicate drought, e.g., see "A" in Fig. B.

The indices with longer-term memories (Table D) are considered to be of greatest potential utility in Texas. To compare these indices, it is useful to examine particular dry spells or wet spells. The Drought of the Fifties, for example, ran from 1950-51 to 1957-58, depending upon

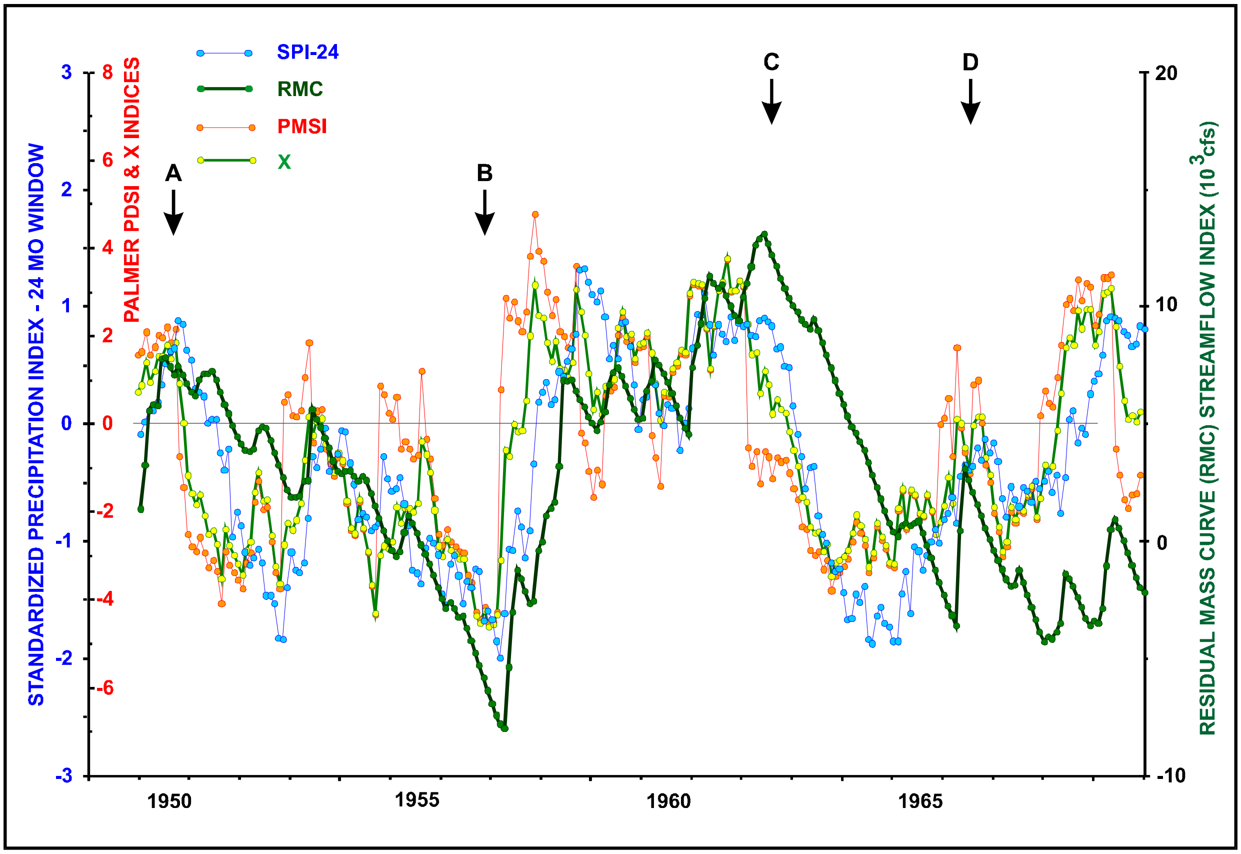

Figure C - Little Cypress Bayou near Jefferson long-term memory indices (see Table D), for 1950-1969

the region of Texas. The 1-year indices generally indicate a return to normal in 1953-54, so would consider this record drought to be made up of two successive droughts. The long-term indices generally give a more realistic depiction of this event as a long, continuous drought of varying intensity. On the Brazos, the Palmer indices correctly display the drought while the SPI-24 delays its onset by about two years. On the Little Cypress, Fig. C, it is the PDSI that behaves erratically, recording two returns to normal during the drought, while the SPI-24 and Palmer X index behave more realistically. However, both the N-index and the SPI-24 are about a year late in detecting the 60's Drought, though the PDSI appears to be more correct. The RMC generally provides an unambiguous detection of the 1950's drought and of the 1960's drought, e.g. "A" to "B" and "C" to "D," resp., in Fig. C, the exception being the upper Brazos at Seymour. In the 1990-2009 period, most prominent are the intense droughts of 2005-06 and 2008-09, and the intense pluvial of late 2007. These are clearly displayed on the Sulphur and Little Cypress by all

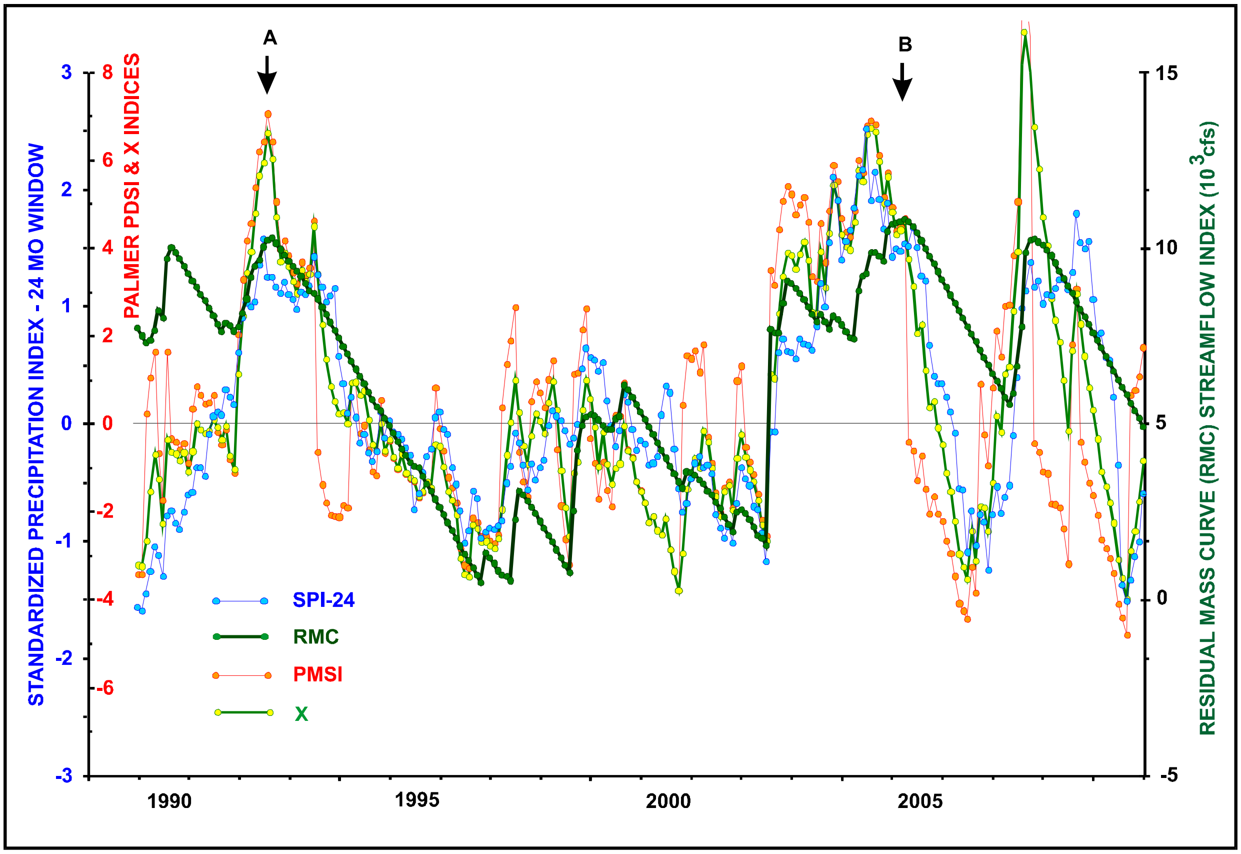

Figure D - Nueces River at Cotulla long-term memory indices (see Table D), for 1990-2009

of the 1-year memory indices and by the long-term indices. The RMC for these gauges, however, displays a protracted drought 2001-07. The Guadalupe, San Antonio, and Nueces gauges show general agreement between the RMC and the rainfall indices, e.g. Fig. D.

The limited resources of this study precluded comprehensive evaluations of the candidate indices, and the results are necessarily provisional. Eight recommendations are offered for future work.

1 It is recommended that the state give further consideration to using six of the candidate indices, namely the PDSI, SPI-12, SPI-24, N index, X index, and RMC.

2 It is recommended that the state further develop the RMC, to identify and explicate its behavior, devise a scaling methodology to normalize the index to a more useful range,

inquire into a decaying-memory formulation, and explore alternative methods for determining the start and termination of drought and pluvial periods.

3 It is recommended that the state compile various data quantifying drought and pluvial *impacts* and use these to "calibrate" the various indices to numerical categories based upon actual drought or pluvial conditions. The lack of a clear relation from indices of surfeit or deficit moisture to quantifiable impacts is considered a major deficiency of all of the indices reviewed, which translates into vague and arbitrary threshold classes, e.g., Tables A – C, above.

4 It is recommended that the data compiled as a result of Recommendation 3 be applied to re-evaluating the protocols for determining the start and end of anomalous moisture periods.

5 It is recommended that if any indices are selected for use by the state that involve equiprobability transformations to a standardized Gaussian (the mathematical method underlying the SPI, SPEI, and several other indices reviewed in this study), the selection of the specific standard skewed distribution to be employed be given detailed study.

6 It is recommended that the use of reservoir contents both as a statewide or regional indicator, and as an index for management of water demand be given more extensive and rigorous study.

7 It is recommended that appropriate thresholds for applicable indices be developed to differentiate the broad (and vague) categories of "wet," "normal," and "dry" used in the Senate Bill 3 environmental flows recommendations for low-flows. This will require the data acquired under Recommendation 3, notably quantifiable impacts from field data.

8 It is recommended that techniques for drought *prediction* in Texas, which were beyond the scope of this project, be given exploratory study. Some possible approaches are suggested.

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Figures

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Tables

Preface

In 2009, the staff of Texas Water Development Board requested that this investigator undertake a modest study to examine water management practices in other countries faced with similar hydroclimatologies to that of Texas, specifically how are variations in surface-water supply quantified and monitored, and what kinds of triggers are used to implement various management measures. Essentially, an international literature review was envisioned, from which a few promising methodologies — if they emerged — might be tested for Texas rivers.

There were several convergent motivations for such a study. The State had recently sustained several short but unusually intense droughts (as well as record runoff events), raising the question of whether such events could be detected more rigorously, and management actions thereby facilitated. This led naturally into the subject of hydrologic indices and their associated triggers. At about the same time, with the passage of Senate Bill 3, the State embarked on a process to formulate standards for flows necessary to attain a desired level of ecological health of its watercourses. Delineation of the resultant "environmental flows" is contingent upon the existing hydroclimatic conditions of the watercourse, which of course varies from drought to flood, but there is no acceptable operational method for specifying these. This likewise led to the subject of hydrologic indices, perhaps along the lines of the method of Indicators of Hydrologic Alterations (Richter et al., 1996), but better suited for Texas. The scope of the study was extended to include hydroclimatic indices, to accommodate the needs of environmental flows.

However, this study remains only preliminary, its main products being the literature summary (presented in Chapter 3), and a selection of candidate indices suggested for further research, whose behavior is displayed in a series of demonstration cases on Texas rivers (presented in Chapter 4). In general, the use of hydroclimatic indices and triggers in water-resource management is not particularly more advanced elsewhere than in this country, but there are several key avenues of research indicated in the recommendations (Chapter 5). This study attempts to be quantitative and specific to Texas. Generally, data are presented in the same units of the original measurements, in order to preserve the significance reported by their originator, though if combined with other data and rounded, the Système International is preferred. Index time series and supporting calculations are detailed in a series of EXCEL® workbooks, available upon request from the TWDB or the author (gward@mail.utexas.edu).

This study benefited greatly from its perceptive, sympathic and patient project officers, to whom this investigator is grateful. The first project officer was Greg Malstaff who conceived the study, and generously shared his insight and his vast store of information, but retired before the study was completed. Mark Wentzel assumed the rôle and exhibited even more patience as this writer slowly ground on. In addition to his usual competency, Dr. Wentzel offered counsel that restrained this writer from putting his foot in it on occasion (though he could not, of course, prevent all such actions). It has been a pleasure to work with both of these project officers.

1. INTRODUCTION

One of the challenges — perhaps the central challenge — of water management in Texas is contending with its extraordinary variation in water supply. No region of the state, from the deserts in the west to the humid forests of the east, is immune from periods of excessive or deficient rainfall. An adequate management preparation for, and an effective operational response to this variation in hydroclimate require some immediate quantitative measure of surface water conditions, not only of the volumes of flow in the streams and rivers and the storage in reservoirs, but how these relate to normal conditions and their likelihood of meeting regional water demands.

Although the need for a metric suitable for assessing present hydroclimate conditions is implicit (and sometimes explicit) in management decisions for releases or storage in the state's reservoirs, the need for a more general metric applicable as well to free-flowing streams was exposed recently in the environmental flows process. The present study is motivated in part by the flow-regime approach of the Texas Instream Flow Program (TIFP), which seeks to determine flows and their variation considered necessary for maintenance of the stream ecosystem. Four components of the hydrograph are specified: two low-flow components (subsistence and baseflow) and two high-flow components (storm-runoff pulses and overbank flows), see TIFP (2002), TCEQ et al. (2008). This flow-component paradigm was embraced by the SB3 process, and a software product was developed under the auspices of the SB3 Science Advisory Committee (SAC) called the Hydrology-based Environmental Flow Regime (HEFR) methodology (SAC, 2011). The standard presentation format of HEFR is a flow-regime matrix, whose rows correspond to flow components and whose columns correspond to months of the year, perhaps aggregated into seasons, see Opdyke (2010) and SAC (2011).

In view of the extremes of Texas hydroclimatology (e.g., SAC, 2004), it is hypothesized that the response of the ecosystem to the streamflow regime (and therefore the flow levels required for maintenance of that ecosystem) will depend upon the general hydroclimatological conditions prevailing at the time. For this reason HEFR includes the capability of specifying three levels of baseflow conditions (each potentially varying with month or season), *viz*. wet, normal, and dry conditions. However, at present no criteria for determining which set of conditions prevails at a point in time have been developed, and this flow-condition classification has proved to be controversial among hydrologists, climatologists and biologists, especially those making up the SB3 Basin and Bay Expert Science Teams (BBESTs).

The overall objective of this project is to survey the literature on hydroclimatological indicators, with emphasis on the international literature for case studies of trigger formulations, then investigate Texas rainfall and streamflow data to explore and contrast metrics of hydrologic variability (such as wet, dry, and normal). This objective lies on the intersection of several vast subdisciplines of meteorology, hydrology, and engineering, and it has been necessary to scrupulously avoid wandering too far into these areas. Some topics had to be avoided entirely, including flooding and its management, hydrograph analysis, economic and agricultural impacts of water excess or shortage, modeling of watersheds and river networks, and water demands.

The intended technical approach is therefore to exploit the considerable work that has already been done in hydroclimatology, especially relating to quantification of aridity and drought, but with special application to the Texas environment. Once meaningful descriptors of hydroclimatological flow conditions are formulated, they are to be investigated for their viability as operational metrics, including the definition of criteria or "triggers" that identify the categories of hydrologic conditions presently manifest in a stream environment, and that will be likely be encountered in the immediate future. Specific technical objectives are as follows:

1. Review the literature from regions around the world that have climates similar to Texas for definition of various strategies for quantification of hydrometeorological and stream-flow conditions, i.e., with respect to definition of indices and/or indicators, especially as they relate to prevailing climate.

2. Apply any of the schemes found in the review for their suitability in Texas. This will be pursued empirically, based upon data records from a selection of test streamflow gauges and watersheds representing various Texas hydroclimatologies. The selection of schemes to be tested is not limited to those in the literature, but can include experimental ideas and concepts formulated in the course of this work.

3. Test these schemes on a few Texas rivers, to determine whether they will be capable of extension to riverine conditions with a large and hydroclimatologically varied watershed.

4. Prepare demonstration cases using these schemes/indices to depict time patterns of river flow, and evaluate their utility as a management device. Particular interest was expressed by the Board in an evaluation of their efficacy to forecast river flow behavior over a forthcoming season.

The limited resources available to this project dictate that this study must be provisional and preliminary, indicating the possible direction of future, more detailed studies, rather than comprehensive. For this reason, the literature survey of (1) is confined to recent work in the primary hydroclimatological journals and readily accessible representatives of the "grey" literature, with emphasis on relevance to the Texas region, and the computational investigations of (2), (3) and (4) are limited to "selected" data records for "demonstration" purposes. For the same reason, this review does not attempt to address all hydroclimatological indices that have appeared in the literature. We also confine this review to indices developed from measurements of hydrometeorological variables, i.e., the indices should be data-based. In particular, indices dependent upon simulations of a hydrological *model* are not considered, mainly because their review would of necessity take us into the morass of evaluating the adequacy of the underlying mathematical model. Several other categories of index are excluded from consideration, as detailed in Section 1.1.2, below. The indices specifically addressed in this review are selected because either they appeared to this investigator to offer some potential for utility in Texas water management, or because they have acquired some currency.

1.1 Indicators and indices

1.1.1 General desiderata

The terminology in the above statements of objectives warrants clarification. In general, an *indicator* is a measured parameter, or a statistic of a set of measurements of the parameter, that serves to track the variation of the state of some complex entity (*indicare*, Latin, "point to," "estimate"). The "complex entity" might be the economy, a geopolitical system, the environment, or some feature of the physical world, such as an ecosystem, or, in the present case, the hydroclimate of a watercourse. Implicit in the above definition, a satisfactory indicator requires the following:

(i) appropriateness, i.e., quantitative representation of a salient aspect of the entity of concern
(ii) routine measurement, with an extant data base encompassing the range in space and time over which variation of the indicator is to be determined
(iii) sensitivity, i.e., responsiveness to variations in the state of the entity on the time-space scales of concern
(iv) communicability, in the sense of depicting the salient aspect on a meaningful numerical scale that is immediately interpretable by the target user or audience.

These qualities have been abstracted and distilled from Tunstall (1979), Gallopín (1996, 1997), Brouwer et al. (2003), Moldan and Dahl (2007), Boulanger (2008), Joumard (2008) and Weilhoefer (2011). Unfortunately, the related terms *indicator* and *index* (as well as less common terms like *index number, index variable, proxy measure*) are frequently used interchangeably. In the present report, an *indicator* is a directly measureable variable or an associated statistic, while an *index* is a numerical aggregation of indicators, for example a weighted average, a multivariate function of same (including ratios), or an empirical orthogonal function (a.k.a. principal component). In this sense, an index subsumes the properties of an indicator (including the above desiderata) but may be more complexly formulated: an indicator is always an index but an index is not necessarily an indicator.

A useful index possesses both simplification and immediacy. These are, of course, relative to the complexity of the phenomena the index represents. Simplification is achieved by reducing the number of variables that must be addressed and by re-formulating these in a functional form with enhanced smoothness and monotonicity. Immediacy involves transforming and rescaling the indicator so that its space-time variation accords with the qualitative properties of the attribute(s) it represents. This may also involve classifying specific ranges of the indicator to correspond to external responses to that attribute(s), such as attributing a deficit of rainfall to various stages of vegetation stress, or to interface with various management actions. Corollary to such classification is the identification of threshold responses and "triggers."

Depending upon the discipline and the intended use of the indicator, other requirements may be imposed. For indicators of natural systems, including the subject of this study, *viz*. hydroclimatology of the Texas environment, these include:

(v) accessibility of the necessary data on a near real-time basis;

(vi) limited computational demands on the user, i.e., either the indicator/index should be readily available or easily calculated from data;

(vii) existence of an extended period of record at sufficient spatial resolution to display normal and extreme variation.

Indices specific to hydroclimatology and/or management of water resources may have additional requirements. In their development of the Streamflow Drought Index for Greece (see Section 3.2.1), Nalbantis and Tsakiris (2009) included the attributes of being easily understood and carry physical meaning, which combine (i) and (iv), above, and being based upon readily available data, which combine (v) and (vi), above. In addition, they specify:

(ix) sensitivity to a wide range of moisture conditions

(x) independence of the geographical area of application

Property (ix) seems to be more than the sensitivity prescribed by (iii) above, by making explicit the need for a hydrometeorological index to encompass the full scope of moisture conditions likely to be encountered in a geographic location. Finally, property (x), that the index be generally applicable anywhere and not be quantified only for a specific location, is important for regional and larger scale climatological studies, but may be debated for application in operational water-resource management that focuses on a specific gauge, lake or watershed. Drought, for example, like politics, is ultimately local, in how it affects individuals, their livelihood and their activities. But drought is also a physical phenomenon whose geographic scale and evolution in time are important clues to understanding its causes. The utility of a single index that characterizes the same relative level of drought, or any other scenario of moisture surfeit or deficit, is clearly evident for analytical purposes. This dictates some means of normalization, in both time and space, of an index.

For analysis of a record of an index, it is convenient to have a companion specification of threshold values, truncation levels, or triggers that categorize moisture conditions (e.g., Dracup et al., 1980; Steinemann, 2003), for purposes of appraising the statistics of occurrence of such conditions and conveniently displaying those results. For operational use of an index, such threshold values are mandatory in being able to determine the prevailing conditions and communicate these to the public. In extremes of flood and drought, such trigger values carry an implication of specific mitigative actions, water restrictions, dam releases, or other management responses. A clear, intuitive index that is objective and readily disseminated to the public has obvious operational advantages.

1.1.2 Hydroclimatological indices

Consonant with the study objectives given above, this study focuses upon indicators and indices of hydroclimatology, mainly the variables of precipitation, evapotranspiration, and streamflow, with an implicit averaging time of weeks to months. The most common such averaging interval is one month, which is often the basis for climatological studies of temperature and rainfall variation, as well as hydrological studies of streamflow variation, water-supply reservoir design, and regional water planning.

The impacts of excess or deficient surface water are varied and broad (Dracup et al., 1980). The range of impacts of drought in particular has motivated a multitude of indicators. In order to keep the scope of the study within reasonable bounds (success in which is belied by the size of the present document), agriculture- or economics-based indicators are excluded from the present analysis.

Similarly, we exclude from the outset a suite of indices that address the extremes of hydroclimatological variables. Examples of these indices include the R5D (highest consecutive 5-day precipitation total in a year), CDD (maximum number in a year of consecutive dry days), R30 (number of days per year with daily precipitation exceeding 3.0 cm), FDD (number of occurrences of eight or more consecutive days without rain per year), peak flows and related statistics, and *dQy* measures of low flow (annual minimum average of a *d*-day sequence of flows with *y* year recurrence). These require a time resolution of daily data or finer, and address events of short duration compared to the climatological time scales (months, seasons, years) that are the objective of the present study. Moreover, the events quantified by these indices can be embedded in a longer period of the opposite general hydroclimatological condition: an intense storm in a sustained drought, or a short sequence of low flows embedded in a longer wet period. Information on these and related indices may be found in Linsley and Franzini (1964), Riggs (1972), Berenbrock (2002), Frich et al. (2002), Ceballos et al. (2004), Kostopoulou and Jones (2005), and Moberg and Jones (2005), among others.

Little mention is made in this review of the U.S. Drought Monitor (Svoboda et al., 2002). This is not so much an index as it is a management service based in part on indices. More specifically, the Drought Monitor maps are developed from current PDSI and SPI data, simulation of soil moisture using NOAA Climate Prediction Center models, USGS provisional streamflow data at selected gauges (as percentiles of normal), and a vegetation health index derived from satellite imagery. This is a useful and worthwhile service of great potential value to operational water management in Texas, but in itself it is not a quantitative index in the sense of Section 1.1, and therefore is beyond the scope of this study.

There have been several reviews of meteorological and hydrological indicators and indices in the literature, many of which focus on a specific hydrological condition such as base flow or drought. Of particular note are the reviews of Wilhite and Glantz (1985), Byun and Wilhite (1999), Heim (2002), Keyantash and Dracup (2002), Cancelliere et al. (2007b), Robeson (2008), Mishra and Singh (2010), and especially Quiring et al. (2007) and Quiring (2009a, 2009b), being sponsored by TWDB and undertaken for the State of Texas.

1.2 An overview of Texas hydroclimatology

Water is among the very few compounds capable of existing stably in all three states—gaseous, liquid and solid—within the ranges of pressure and temperature on the earth's surface. Its conversion from one state to another facilitates its movement on the earth and the associated absorption, transfer and release of energy are essential to the mechanics of the atmosphere, oceans, and land surfaces. Hydroclimatology in general refers to the characterization of the presence of water substance in a space-time domain. The characterization is typically, but not exclusively, statistical, and the space-time domain is usually, but not exclusively, confined to some geographical region within a specified time period, often, but not exclusively, delimited by the period of record of observations.

The large-scale climate of Texas is dictated by four principal controls: (1) the belt of westerlies, (2) the onshore flow from the Gulf of Mexico, driven by the trade winds, (3) the state's physiography, comprised of the Rocky Mountain massif to the west and a low-relief plain to the east sloping down to the Gulf, and (4) solar radiation at the surface. The descent of the westerlies to the plain in the lee of the mountains results in a warm, dry airmass, the "rain shadow" of the Rockies. The southeasterly onshore flow from the Gulf is tropical air, warm, humid, and typically convectively unstable through a considerable depth, and is the primary source of water vapor to the state. (As noted in Ward, 2005, these same controls govern the large-scale climatology of the contiguous United States. Because of their geographical convergence with distance south on the continent, Texas in many respects mirrors the hydroclimatology of the continent east of the Rockies.)

Convection is stimulated by disturbances in either airstream, i.e., the westerlies and the trades, and their interaction, and is often enhanced by the terrain. The relative influence of the westerlies and trades, and their respective disturbances, varies seasonally with the movement of the climatological equator. Precipitation is predominantly due to deep convection, and is almost entirely in the form of rainfall, which increases markedly from the arid west to the humid east, averaging a factor of six to seven across the state. This pattern of rainfall is exemplified by the isohyets of Figure 1, which typically lie parallel to meridians (and parallel to the axis of the

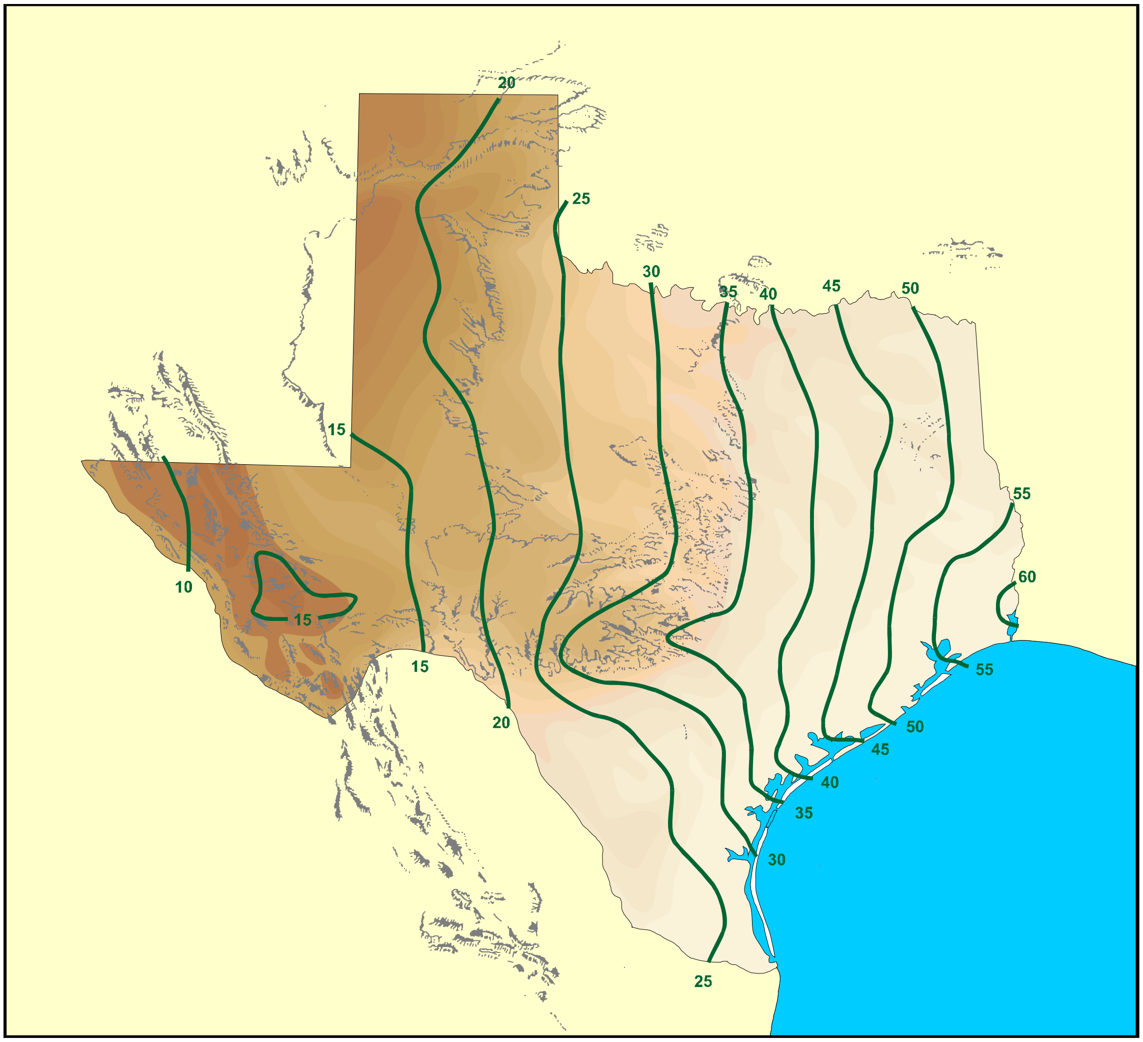

Figure 1 - Texas 1971-2000 normal annual isohyets (inches/year)

Rocky Mountains), except for obvious orographic perturbations. Its deep-convection origin implies that rainfall occurs in brief bursts sparsely distributed in time. The magnitudes and time-clustering of bursts increases from west to east across the state and is generally a function of the seasonal prevalence of synoptic-scale disturbances. The general seasonality of monthly rainfall is indicated by the sketch of Figure 2, showing the distribution of monthly maxima by the seasons in which they occur. In winter, there is more than a forty percent increase from north to south in solar radiation over the latitudinal extent of Texas. The resulting mean isotherms lie parallel to lines of latitude and there is a 25-30°F temperature difference across the state.

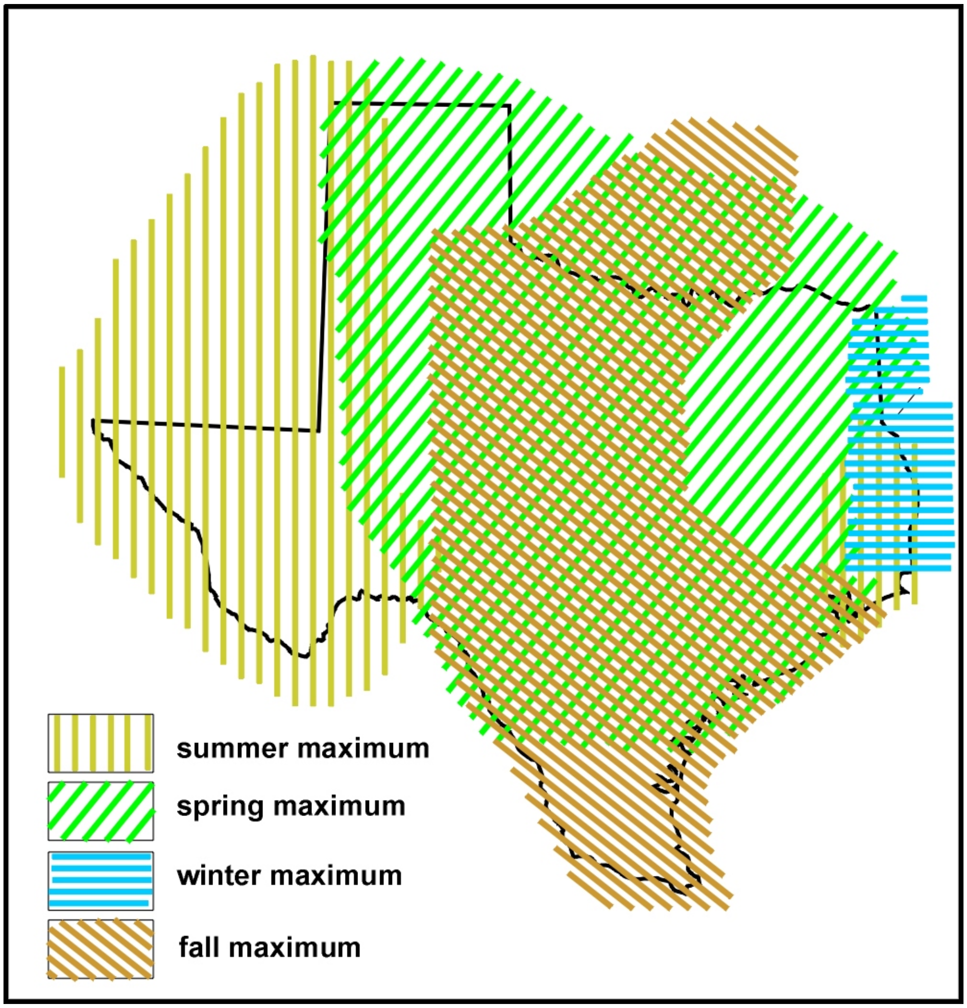

Figure 2 - Seasonality of monthly rainfall (Ward, 2005)

The ultimate source of water in a stream is precipitation, but the time lag between water impingent on the watershed surface as rainfall and its appearance in the stream may range from minutes to centuries depending upon its trajectory. After an initial debit of rainfall due to surface and vegetation interception, and to filling of intergranular spaces in the near-surface soil, water begins to infiltrate into the soil. If the rate of precipitation exceeds that of infiltration, the surplus ponds on the surface, collecting in depressions. As rainfall is maintained, ponding continues until hydraulic continuity is established and the water can flow downgradient as runoff. This flows into the surface drainage network organized into successively larger conveyances, from rivulets and runnels to streams and rivers.

The short-term influx in the stream is therefore the direct response to a precipitation event, in Texas usually a thunderstorm, but lagged and extended in time due to the complex of transports from the watershed surface through the drainage network to the stream. Additional streamflow follows transport pathways in soil (vadose zone) as interflow and from groundwater (phreatic zone) as influent when the stream channel cuts the aquifer. An important component of the stream interflow arises from the soil and bedrock zone surrounding the stream. This zone may merge with the surrounding vadose zone of the watershed, in which case it offers an avenue of (usually slow) flow from uplands to the vicinity of the stream. This zone may also connect with more permanent aquifers. Temporary "bank" storage is forced into this zone when a storm hydrograph exceeds the elevation of the water table, which then acts as a source of streamflow after the storm hydrograph recedes.

Unfortunately, the relation between rainfall and streamflow due to the combination of all of these processes is often ambiguous, in at least two respects. The first is the difficulty in separating the immediate runoff from a storm from the longer-term flow in the stream deriving from extended but temporary storage of the rainfall. The convention is to refer to the former as storm flow (or quick flow, or direct flow), and the latter as base flow (e.g., Dingman, 2002). The second is in that different combinations of watershed and interflow processes may yield the same streamflow response to rainfall. This means that an analysis of data on streamflow and rainfall alone cannot resolve the underlying physical processes, a situation referred to as the "equifinality" of multiple solutions (Beven, 1993, 2006; also see, e.g., Jakeman et al., 1990; Jakeman and Hornberger, 1993; van Dijk 2010).

An example of storm-pulse hydrographs is shown in Figure 3, at the Goliad gauge on the San Antonio. The storm pulses are superposed on a much smaller sustained flow, and are clustered seasonally (cf. Fig. 2). As stated above, because the focus of this study is hydroclimatology on longer time scales (months or greater), details of the storm runoff process and the analysis of storm hydrographs (or "pulses") are excluded from consideration, other than inclusion in the longer-term average flows. However, the notion of base flow is potentially important, as this is one of the key streamflow components identified by TIFP and HEFR. There are a number of "base-flow separation" procedures available, many embodied in computer programs.

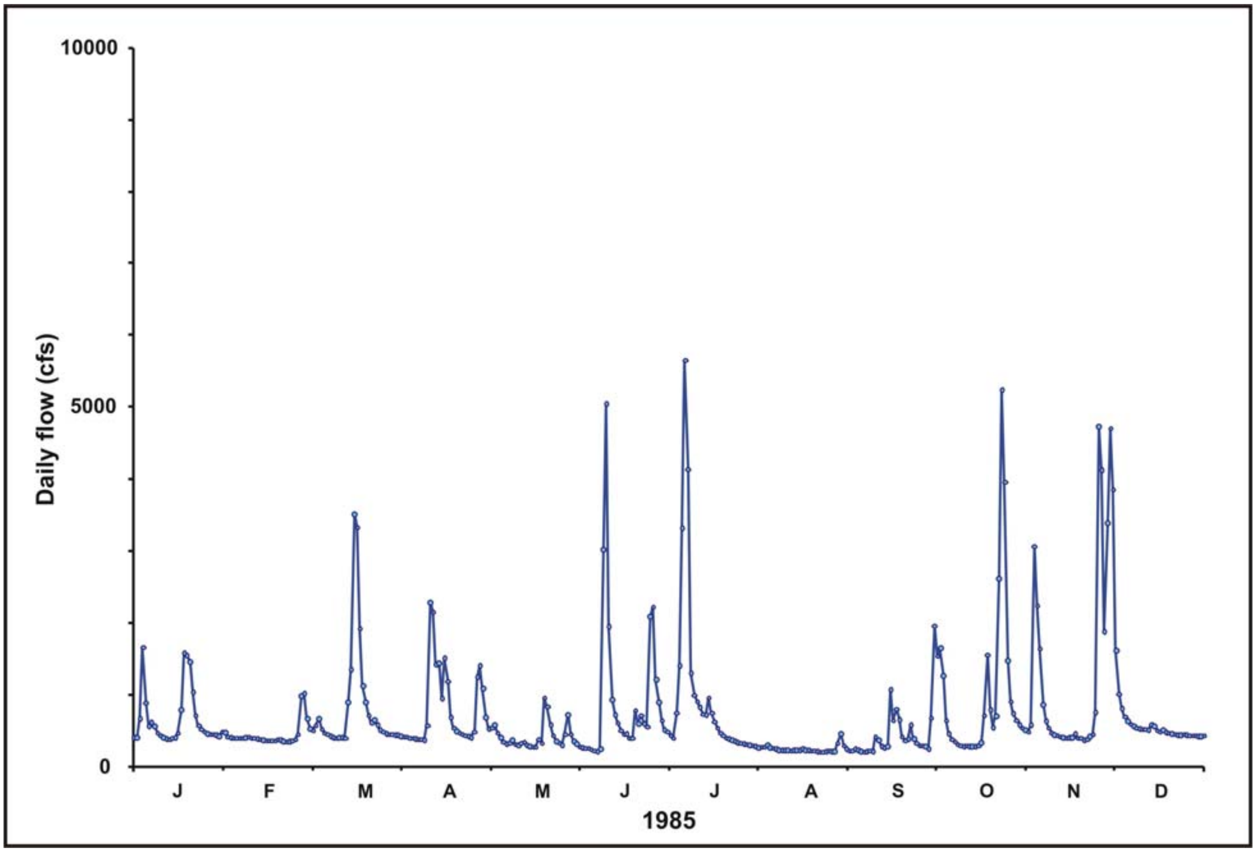

Figure 3 - Daily flows San Antonio River at Goliad, 1985

Mathematical methods for base-flow separation generally focus on the recession rate of the hydrograph, and range from simple graphical fits to digital signal processing (see, e.g., Hall, 1969, Nathan and McMahon, 1990, Arnold et al., 1995). Whatever the underlying conceptual model, these methods do not differentiate between flows from distant reaches of the stream network, various vadose-zone sources of interflow, including bank storage, and groundwater discharge into the stream channel. Rather, they effectively interpret "base flow" to mean low-magnitude, relatively steady flow that remains after the transient of a flow pulse has died out. This is the meaning of the term employed in the present study. We note that, while the low, steady character of the flow is its defining attribute, it does not follow that base flow has only one source or only one magnitude. Indeed, it is considered to vary in response to long-term hydrometeorological conditions.

Table 1
Mean annual water budget of Texas, 1971-2000 normal, Bm^3/yr
see Ward (2005, 2011)

	State water budget		
Input from atmosphere		*Disposition of surface water runoff*	
Precipitation	514.0	Inflow from out of state	14.4
Evapotranspiration*	– 440.1	Diversions net of returns††	-3.1
Net P – E	73.9	Outflow to other states	-15.7
Disposition of annual water crop		Flow to sea	46.3
Lake evaporation	8.2		
Groundwater infiltration †	6.8		
Surface runoff**	58.9		
	Human uses		
Human consumption (net††)		*Sources for human water demand*	
Agriculture	11.6	Groundwater	12.4
Municipal & industrial	4.0	Surface water (net††)	3.1

* Estimated as precipitation minus the sum of runoff, groundwater infiltration and lake evaporation
** USGS regional regressions (Lanning-Rush, 2000)
† TWDB data (ca. 1980 with 1951-80 normal)
†† Net of all returns including groundwater sources

A general water budget for the State of Texas is presented in Table 1. This is a statewide, long-term average budget. Both water supply and water demand vary strongly with geography (as suggested above), as well as in time, which is the basic challenge of water management in the state. More detail on the geographic variability of the budget, and budgets for critical periods, i.e. drought, as well as more detail on water uses, may be found in the citations. The separate entries of Table 1 are based upon data (with the exception of evapotranspiration, which is estimated), therefore the components do not exactly balance due to uncertainty in the individual measurements. However, these provide the general magnitudes of the key components of the water budget. Probably, the most important observation to be made is that the total water influx to the state is the difference between two large, nearly equal quantities, precipitation and evapotranspiration. Relatively small variations in these two quantities result in large variations in their difference.

1.3 Storage and memory

Most of the time, in most places, precipitation does not occur. In Texas, in particular, rainfall is produced mainly by thunderstorms whose space-time distribution is sparse. The central feature of the hydrological cycle that allows nearly continuous access to water by all living things is the ubiquitous opportunity for storage. Water is stored in surface depressions, in the network of stream channels, in layers of the soil, and in aquifers. The retention in the various types of hydrological storage ranges from minutes to millennia, over nine orders of magnitude. Humans have improved the water-supply abilities of the hydrological cycle for their own purposes by creating additional, strategically situated modes of storage.

The flow in a stream channel can be viewed as the time-series output to the input time series of precipitation on the watershed, in which the watershed acts as the signal processor. Probably the most fundamental such example is the linear response of a small watershed to a brief intense burst of rainfall, in which the response is a rise to a peak during the rainfall event followed by an exponential decay $\exp\{-kt\}$. A measure of the rate of decay is the time constant or time response $T \equiv 1/k$. This time response is due to the temporary storage of the input rainfall in the watershed (and can be modeled by a linear reservoir, e.g. Dingman, 2002). For larger and nonhomogeneous watersheds, for which the rainfall input may be both spatially and temporally variable, the stream flow response is lagged and protracted, dependent upon the network of stream channels and subsurface permeability upstream from the gauge, among other things. Despite this complexity, the decay after the peak flow remains a useful indicator of the watershed response at that gauge.

Often an index $z(t)$ can be depicted as a function of time obtained from the time series of a hydrometeorological variable $x(t)$ by a convolution

$$z_i = h*x \equiv \sum_{k=1}^{N} h_k x_{i-k} = \sum_{k=1}^{N} h_{i-k} x_k \qquad (1)$$

(This is a discretization of the expression for continuous functions $z(t) = x(t) * h(t) \equiv$

$\int_0^T h(\tau)\, x(t-\tau)\, \mathrm{d}\tau$ in which $x(t)$ is defined over the finite range $t = 0$ to T.) The kernel of the convolution h_k can offer insight into the nature of the index as a signal processor. The index transformation is in effect a filter, and we will find it useful to try to expose the underlying kernel to determine the characteristics of the filter. Alternatively, we will estimate the underlying time constant by either a best-fit analysis of an exponential, or by the settling time response to a unit impulse (which is about three time constants). If z is linear, the kernel h in (1) is the response of z to a unit impulse in x. It is therefore an expression of the memory of the function $z(t)$. If $h(t)$ decays quickly to zero, then the function $z(t)$ has a short memory and is responsive to the most recent values of $x(t)$. In contrast, a slowly decaying or even constant $h(t)$ implies a sluggish response and a long memory to past values of $x(t)$.

There are therefore analogs between the behavior of a real watershed and the abstract filter that is implicit in a hydrometeorological index. Both the index and the real watershed act as signal processors of the rainfall time series, and both have memory that determines the nature of their responses. Though the same terminology may be applied to either, as indicated above, a clear distinction must be made between the two. The behavior of the watershed is physical, and derives from the process-trajectories of water from a rainfall event. Mathematical filters, such as a weighted moving average, are employed for data processing purposes, to suppress certain features of the response to rainfall (say) and amplify others. Most of the indices considered in this study have properties that are determined by a desired application, which is not to mimic a real hydrometeorological variable but rather to selectively expose certain aspects of its behavior. (Apart from the above terminology and this cautionary note, signal processing is not extensively applied in this review. Should detailed information be sought, Bendat and Piersol, 1971, Cadzow, 1973, and Karl, 1989, are recommended.)

2. HYDROCLIMATOLOGICAL DATA

This report of course focuses upon water in its liquid state, especially its concentration in quantities necessary to support or impact human activities, with specific application to Texas. An essential prerequisite for the employment of a candidate indicator is the availability of data, both in near real-time and as a historical record. The purpose of this chapter is to summarize the types and properties of data suitable for use in formulating and applying various indicators in Texas water management. Specific sources for data are generally already known to the TWDB or are summarized in other reports (see especially Quiring et al., 2007; Lackey et al., 2012), so are not presented here. Rather the emphasis here is on the space-time distribution of data and sources of error, and their subsequent processing, an emphasis informed by the history of measurement and data collection.

Routine measurements quantify either the flux (or transport) of moving water, or the volume of water contained in storage. Perhaps the most fundamental such measurement is the rate of rainfall, as measured by the depth of water collected after passing through a horizontal orifice over some interval of time. The collection rate of depth per unit time is really the volume flux through the area of the orifice. More generally, *precipitation* is the measured parameter, reported in equivalent depth of water, which includes the frozen forms. For Texas hydroclimatology, however, *rainfall* is predominant, and the two terms are used interchangeably hereafter. The rate of transport of flowing water in a stream channel, i.e., streamflow, is the variable of central importance to water resources planning and management. In most practical situations, the elevation of the water surface parameterizes flow in the channel, through relations from open-channel hydraulics, and also serves as an indicator variable. Though measurement and data collection are not addressed here, the volume of storage on the watershed most amenable to use in water management is the contents of reservoirs, for which the surface elevation is an indicator.

2.1 Precipitation

2.1.1 Data collection programs

Collection of meteorological data from a network of stations has historically taken two different strategies. The first is to equip a large number of amateur volunteer observers with simple instruments, who then make daily measurements, which are compiled at some regular interval of time by the observers and mailed to a central collating and analysis facility. This is referred to as a "cooperative" program, and its purpose is strictly climatological. The second is to implement a (usually broader) suite of regular meteorological measurements, performed by professionals (but not necessarily meteorologists), that are communicated *immediately* to the central facility for analysis, referred to as a "synoptic observation" program. The objectives of this strategy are to track the development of weather on as nearly a real-time basis as possible, and to provide a basis for weather forecasting.

Cooperative programs have been operated in the United States since early in the nineteenth century (e.g., Dupigny-Giroux et al., 2007), though generally on a small scale, such as the U.S. Army Medical Department observations at Army posts initiated in 1819, the State University of New York network established in 1825, and the Pennsylvania network in 1837. A nationwide cooperative program was created under the auspices of the Smithsonian Institution in 1847, eventually growing to over 600 observers. A synoptic program had to await the invention of the telegraph, which provided the capability for rapid transmission of data. Again under Smithsonian sponsorship, a synoptic network of some twenty telegraph operators was established in 1849 for the purpose of national weather status reports. These were incorporated into a rudimentary (by modern standards) forecasting service, eventually disseminated to a few major newspapers. Both Smithsonian data-collection enterprises were decimated by the Civil War.

Since the Civil War, the task of collection and archiving of meteorological data, including rainfall, has mainly devolved upon the federal government, beginning with the Signal Service of the Army in 1870 (Division of Telegrams and Reports for the Benefit of Commerce and Agriculture, see Hazen, 1884), and in 1891 succeeded by the U.S. Weather Bureau (USWB) in the Department of Agriculture (USDA), see Weber (1922). The synoptic program of the Signal Service was manned entirely by military personnel, and ultimately grew to about 140 stations (plus about a dozen in Canada). In 1874, at direction of Congress, the remnants of the Smithsonian cooperative program were transferred to the Signal Service, and were augmented by reports from medical doctors at army posts. By the early 1880's this cooperative (or "climatic") program comprised over 450 observers (Hazen, 1884). The program began to be reduced in 1881 and was suspended entirely in 1884. The stated reason was that the advent of "self-registering" instruments eliminated the need for cooperative observers (Weber, 1922). It was in fact an initiative of the Chief Signal Officer to encourage the states to take up climatological data collection. The idea of organizing state weather services was suggested to him by Dunwoody in 1881, and the states responded positively (Dunwoody, 1896). (In 1887, the federal cooperative program began to be revived somewhat, see Whitnah, 1961.) After 1891, under the operation of the USDA Weather Bureau, the state climatological programs were integrated back into the federal agency.*

The archiving and climatological reduction of data were viewed as ancillary to the main objective of preparing and disseminating weather forecasts, so historically were undertaken by a separate department within the weather service. In 1887, finally, the Signal Corps (as it became known) created a records division to consolidate all cooperative data. With the transfer of weather service functions to the Weather Bureau in 1891, the new agency assumed responsibility for the operation of the cooperative stations, including the state programs (see above), and reduction, analysis and archiving of the data collected. At the time, the number of first- and second-order stations† totaled 172, the federal volunteer observers numbered roughly 400, including about 160 surgeon officers, and the state cooperative programs about 2000 in total

* Publication of the monthly average data for the state cooperative programs began in the January 1887 *Monthly Weather Review* (a publication originated by the Signal Service) and continued until October 1895.

† A first-order station is permanently staffed with observers, and measurements are made around-the-clock of a comprehensive suite of atmospheric variables. In the nineteenth century, second-order stations operated the same way, but with reduced suite of measurements and concomitantly reduced staff. Third-order stations employed paid observers but reported only min/max temperatures and precipitation, like the volunteer cooperative stations. Whitnah (1961) reports that in 1890 the respective numbers of each type of station were 26, 118 and 34. The source for the above 1891 total of 172 is also Whitnah (1961). Now, a second-order station is professionally staffed by another agency, and the observer(s) trained by the National Weather Service (Shea et al., 1994). Most of these are FAA stations.

(Whitnah, 1961). By 1897, there were over 3000 cooperative observers (Moore, 1898). In 1906 this activity was organized into the Climatological Service, later re-named the Climatological Division. By the early 1920's over 4500 cooperative observers were in the network (Weber, 1922). Presently, the data management rôle is carried out by the National Climatic Data Center (NCDC) within the National Oceanic and Atmospheric Administration (NOAA). The number of active cooperative observers is approximately 8,000 (Shea et al., 1994).

In Texas, there have been historically over 8,500 stations, of which around 630 are currently active. Many of these stations, both historical and active, represent slight shifts in position of earlier stations, so the geographical density is exaggerated. The general distribution of the current station network is shown in Figure 4.

2.1.2 Data aggregation

Climatological summaries of the NCDC are based on daily meteorological observations. There are presently two sources of such data, professionally staffed stations including the National Weather Service (NWS) first-order stations and second-order stations, and the Cooperative Observer Network, as summarized above. (Traditionally, the NCDC has compiled these into TD-3200, the "Surface Land Daily Cooperative Summary of the Day.") From these data the basic time aggregations for presentation are monthly averages and annual averages.

Climatological statistics requires not only averaging in time but some sort of aggregation in space. In part, this is to depict the larger scale, regional variation in climate. It is also a means of compensating for the geographical sparsity of data collection, especially for rainfall. The average density of cooperative stations in Texas is at most one station per 1000 km^2 (i.e., an average separation of at least 33 kms, or 20 mis). This should be compared to the "footprint" of the convective storm system giving rise to the precipitation. For deep-convecting systems, precipitation is concentrated in single cells or clusters of cells. At the low end of the size spectrum, these cells are airmass thunderstorms, typical of summer in most of Texas. On average, these single cells are about 30 km^2 in area, and rarely smaller than 20 km^2 (e.g., Morin

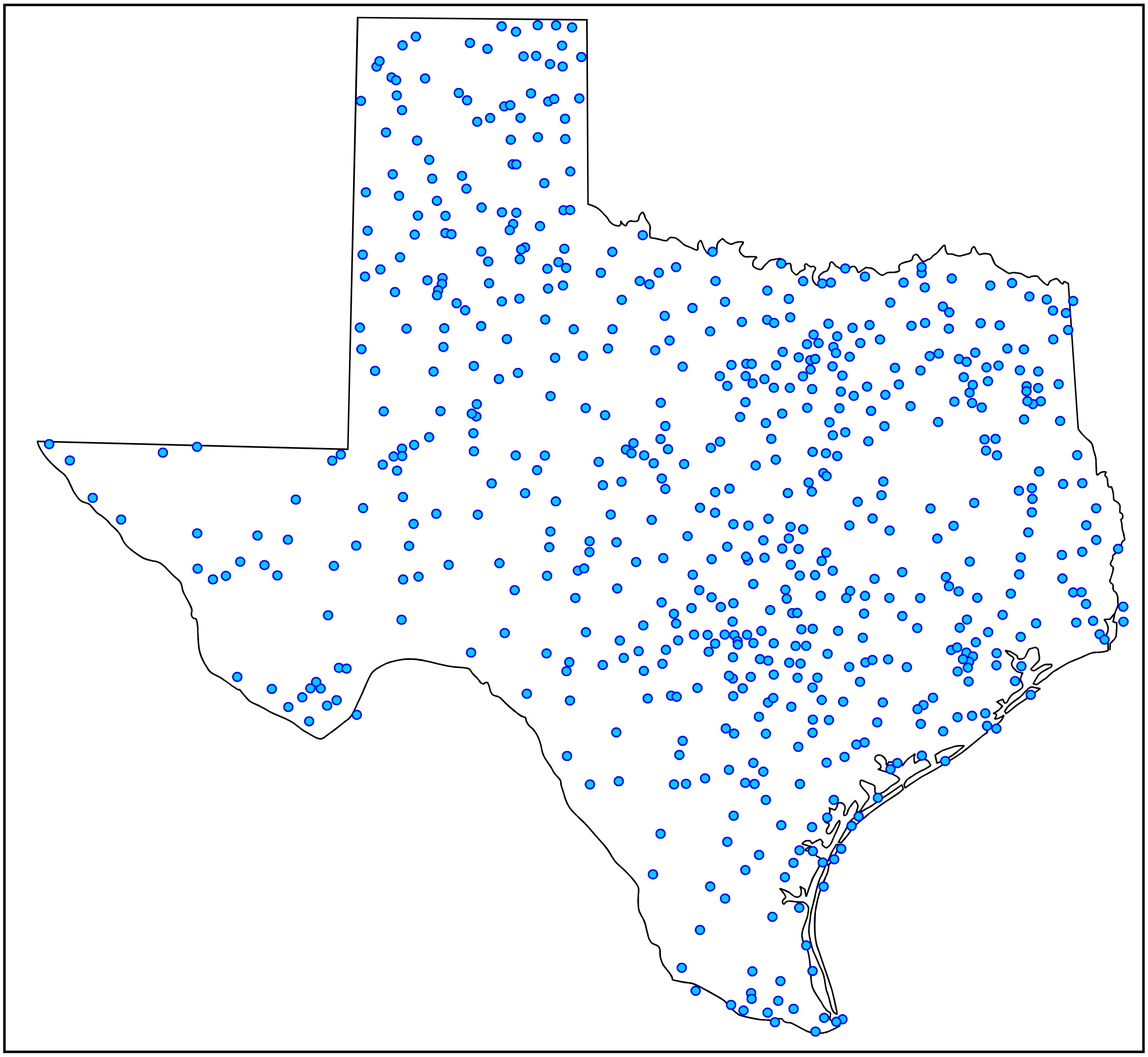

Figure 4 - Approximate locations of active NCDC cooperative stations in Texas

et al., 2006). At the other end of the spectrum, "supercell" thunderstorms have precipitation areas ranging up to the order of 10^4 km^2 (e.g., Smith et al., 2001, Bluestein, 2009). There is a high probability that an airmass cell will be routinely missed by the sampling network, while a supercell storm may be sampled by one or two stations. For rainfall originating from these types of storms, the sparsity of sampling will induce a scatter in the reported rainfall. Aggregating in space may average out this scatter over a protracted time period, though the representativeness of the rainfall data for any single event is questionable. The sampling density is better equipped to sample rainfall from larger "synoptic-scale" weather systems. Mesoscale convective complexes are clusters of cells exceeding specified thresholds of size and intensity, and which are longer lived than single-cell storms. These MCCs have areas of significant precipitation on the order of 10^6 km^2 (Kane et al., 1987), which on average would be sampled by about 10 stations. At the largest scale are the precipitation patterns associated with synoptic systems, such as cyclonic storms, frontal passages and squall lines, whose precipitation areas can range up to 10^8 km^2 or more. It is, of course, these systems that the cooperative network is best suited to monitor, though it still may imperfectly sample the rainfall maxima.

Ideally, the spatial units for aggregation should be regions of quasi-homogeneous climate. However, in the U.S., and in Texas in particular, other considerations have influenced their specification. In the early operation of the Signal Service and later the Weather Bureau, there were so few reporting stations that data were simply summarized by station. With the formation of the Climatological Service in 1906, later the Climatological Division, within the USWB, data began to be organized by states, or groups of states. In 1909, the United States was divided into twelve large topographic "districts" (Guttman and Quayle, 1996), each a principal drainage area of the continent, whose boundaries generally cut across states. (Four of these were major subdrainages of the Mississippi basin, of which the southernmost encompassed the Red River basin.) In the period 1910-1914, the district organization was phased out and replaced by a hodge-podge system of 106 "climatology" sections, whose boundaries were mainly state lines and geometric subdivisions. (Guttman and Quayle, 1996, state that these boundaries were based upon "mailing practicalities.") By 1922, these had been replaced by 42 sections, most of which were single states, with a few combining two or more states (Weber, 1922). With slight modifications, this system was maintained through the 1940's. In 1949, these sections were subdivided into divisions that coincided with the USDA Crop Reporting Districts (Guttman and Quayle, 1996).

In the 1950's the present climatic division system was established. Each state is subdivided into one to ten climatic "divisions," whose boundaries generally follow county lines, but whose shapes and distribution reflect a mixture of terrain, drainage, airmass exposure, population, and

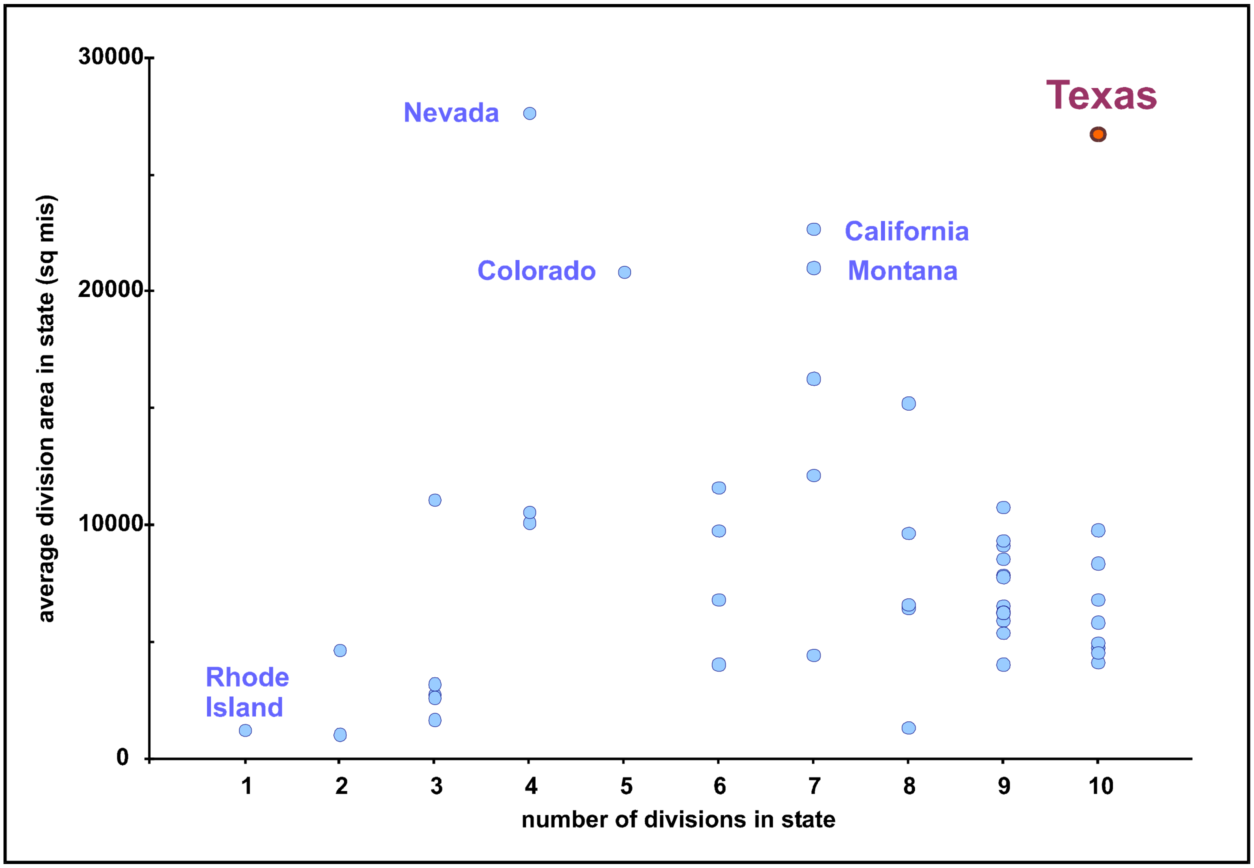

Figure 5 - Distribution of average area of climatic divisions by state versus number of divisions in state

human activity. More than half of the states have eight to ten divisions, the number generally, but not uniformly, increasing with the size of the state, which would imply that the average division area would likewise generally increase with the size of the state, see Figure 5. The NCDC climatic divisions for Texas are listed in Table 2, along with their surface areas, and are mapped in Figure 6. It is evident that in Texas the climatic division boundaries follow county lines.

With each revision of the national system for geographical aggregation of meteorological data, the archived data could be re-distributed in the revised system to construct a historical record for each new geographical unit. Guttman and Quayle (1996) describe the construction of such a record for the present climatic divisions extending back to 1895. For the period prior to 1931,

Table 2
National Climatic Data Center Climatic Divisions of Texas (see Figure 6)

Division number	*Name*	*Area*		*number of stations*	
		sq mis	*10^3 sq kms*	*historical*	*active*
1	High Plains	39406	102.1	938	89
2	Low Rolling Plains	25558	66.2	758	56
3	North Central	39005	101.1	1955	122
4	East Texas	33204	86.0	1241	79
5	Trans Pecos	36171	93.7	598	49
6	Edwards Plateau	35289	91.4	1144	76
7	South Central	21467	55.6	969	81
8	Upper Coast	12217	31.7	552	47
9	South	21949	56.9	450	33
10	Lower Valley	3074	8.0	185	17

the spatial coverage is often unsatisfactory, or missing entirely, in which case the division averages had to be estimated by statistical regression from statewide averages. This data, i.e., monthly temperature, precipitation and drought indices by climatic division, are issued by NCDC in its TD-9640 file, updated monthly.

A major revision of the geographical basis for analyzing and reporting climatological variation in the contiguous United States is presently underway at NCDC. This will replace the present climatic division spatial organization with a rectilinearly-gridded divisional data set. The data themselves will be based on the Global Historical Climatological Network – Daily (see, e.g., Menne et al., 2009, 2012), which incorporates the same historical network of measurements described above, but employs state-of-the-art quality-assurance protocols together with sophisticated methods of bias elimination and interpolation. It is unclear whether climatological products will continue to be produced for the present divisions, but we assume so.

There have also become available other spatial aggregations of the Cooperative Observer data sets. These are of value mainly for research purposes, as they are not routinely updated with current data. An example is that described by Maurer et al. (2002). This re-sorts the 1950-99

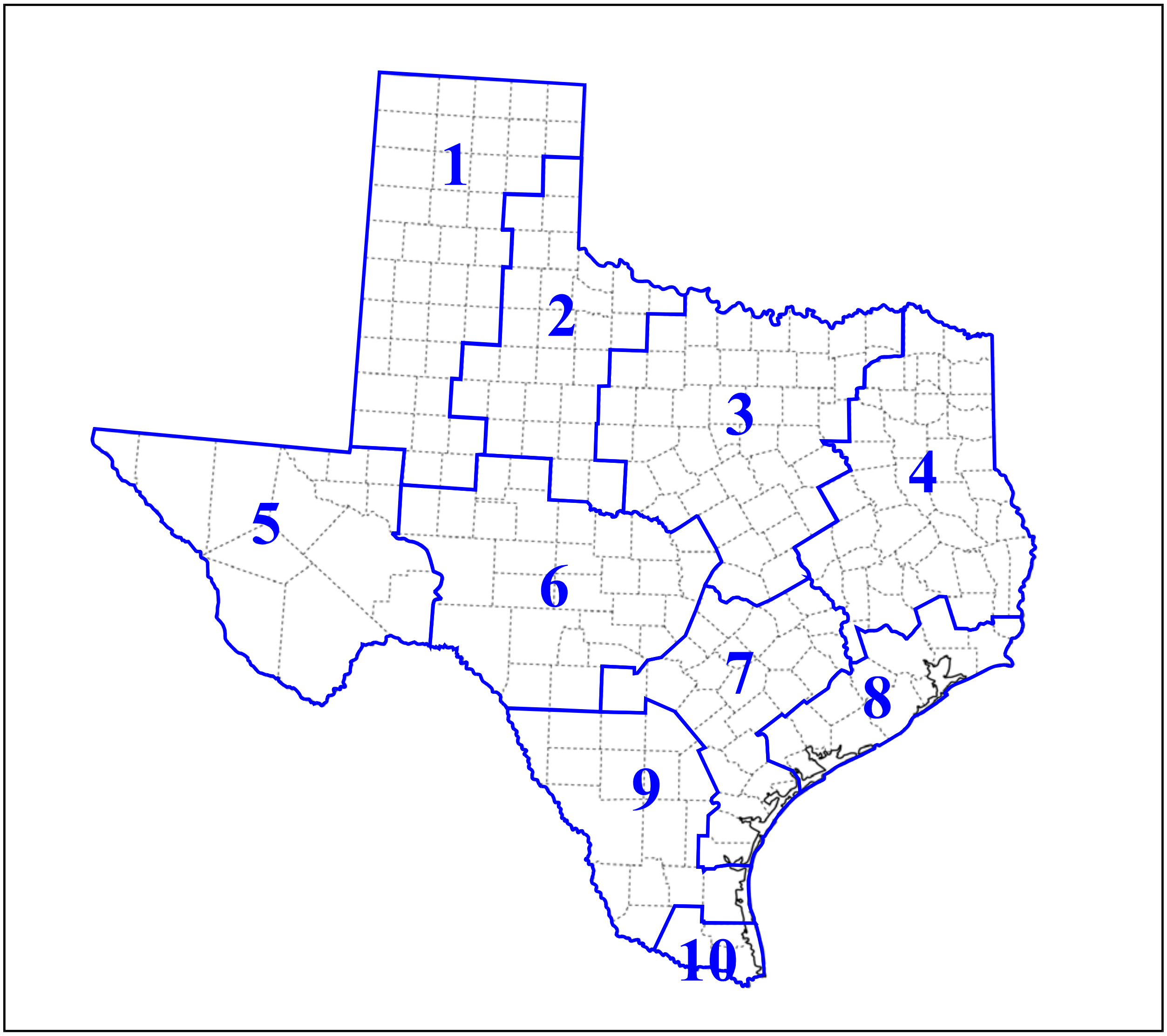

Figure 6 - Climatic Divisions of NCDC in Texas

daily data into a 1/8º grid and combines it with additional model-constrained flux data. Another example is the parameter-elevation regression on independent slopes model (PRISM) of Oregon State University, which combines the Cooperative Data of NCDC with data sources of other agencies and newly digitized historical data, carries out an extensive QA of the data, employs a digital elevation model to quantify the effects of topography, proximity to water, etc., on the internal consistency of the data and as a basis for data infilling, and finally interpolates station data to a 4-km grid over the continental U.S. of monthly precipitation and temperature. The original data record was created from 1895 to 1997, and later extended to 2003. Details are given by Daly et al. (2002) and citations therein (available from NCDC). Rupp et al. (2010) performed a gridding of the Global Historical Climatological Network – Daily for the State of Texas and adjacent regions in neighboring states. They interpolated the 1901-2000 data to a 0.2 x 0.2 grid and carried out an extensive error analysis.

2.2 Streamflow

2.2.1 Data collection programs

The measurement of the flow conditions of a stream is an ancient enterprise. There are fundamentally two types of measurement. The first is the stage of the stream, i.e., the elevation of the water surface, either directly by the intersection of the water surface with a known elevation standard, e.g., a fixed graduated staff, or indirectly through the elevation of a high-water mark. Because the measurement is simple, this kind of data dates back thousands of years. The record of stage elevations of the Nile at Cairo extends nearly continuously from the seventh century to the present (Popper, 1951). In Texas, virtually all reported river-condition data in the nineteenth century are stage observations (e.g., Baker, 1875; Taylor, 1930).

The second type of measurement is flow in the stream channel. This had to await basic advances in hydraulics and instrumentation, and emerged from the Renaissance, notably the work of Leonardo, Galileo, Castelli and Guglielmini, see Levi (1995). In the U.S., the seminal work was the 1851-1860 flow gauging of the Mississippi and its tributaries, undertaken by the Corps of Engineers under the command of Capt. A. A. Humphreys, with much of the hydrometric work directed by Prof. C. G. Forshey, see Humphreys and Abbot (1861). This study, by its scale, methodology, and precision, laid the groundwork for measuring river discharge by the velocity-area method.* The next great step in stream-gauging was taken by Maj. J. W. Powell upon the creation of the Irrigation Bureau in 1888, in convening a decade of engineers at Embudo, NM to study and evaluate methods for stream gauging. This became an extended field seminar under the leadership of F. H. Newell (Frazier and Heckler, 1972). The participants camped on the banks of the Rio Grande for months, implementing various alternative measurement techniques including the use of current meters (which had begun to appear as various *ad hoc* designs in the previous two decades, e.g., Frazier, 1964) and making excursions to other gauging sites in the Southwest. Over the next several years, after the demise of the short-lived Irrigation Bureau, the

* Current meters were not yet generally available. Instead, the vertical-mean current was measured by tracking the movement of two drogues tethered together, one at the surface and the other ballasted to remain just above the river bed. The depths of the drogues were based on hydraulic studies of current profiles.

protocols and methodologies for flow gauging evolved in fits and starts within the Geological Survey (Follansbee, 1994). The successive refinement of these protocols is documented in a series of USGS field manuals (e.g., Newell 1901b; Murphy et al., 1906; see also Hoyt and Grover, 1907).

In the late nineteenth century, starting with the station at Embudo, USGS slowly built a network of gauging stations, at first mainly in the West. In Texas, the first gauging station was established on the Rio Grande at El Paso in 1889 and operated by USGS until the second half of 1931, when the International Boundary Commission (IBC) assumed its operation (and publication of the data, IBC, 1931).* In 1897, Prof. T. U. Taylor, who had a long interest in stream gauging, was retained by the USGS on a per-diem appointment as Hydrographer (Newell, 1899). He selected stations on the Brazos, Colorado, Guadalupe, and Pecos, added a station on the Trinity the following year, and began a series of special studies on Texas water that appeared later as water-supply papers of USGS (e.g., Newell, 1899, 1900, 1901a). The early measurements were of stage only, which were later converted to discharge once the stations were rated. In addition to the El Paso station (whose daily flows for 1893-96 appear to be missing), three of Prof. Taylor's original stations, the Brazos at Waco (USGS 08096500), Trinity at Dallas (08057000), and Colorado at Austin (08158000), have the longest continuous record in the state, dating back to 1898. Just after the turn of the twentieth century, sixteen gauges were operative in Texas, Table 3, a few more being established within the first decade. In 1915, the Texas Board of Water Engineers was funded for stream gauging and began a long-standing cooperative agreement with the USGS that has enabled the development of a gauging network in the state (Ellsworth, 1939, Breeding et al., 1964). By 1916, there were 35 operating stations, and the number grew to exceed 250 by the mid-1920's. Unfortunately, the dependency on state funds led to rather extreme fluctuations in the number of stations over time, and significant gaps in the records of many gauges.

River stage observations were also carried out by the Signal Service since its inception in 1870,

* This apparently authoritative information, which is repeated in other sources, cannot be reconciled with the statement of Grover (1917) that, "The stations on the Rio Grande from El Paso to Eagle Pass, Tex., were maintained and the records furnished by the Commission for the Equitable Distribution of the Waters of the Rio Grande." Someone should look into this.

Table 3
Gauging stations operative in Texas in 1902 (Walcott, 1903)*

U.S. Geological Survey			
Trinity	Riverside	Guadalupe	Cuero
Brazos	Richmond	Pecos	Moorhead
Brazos	Waco	Pecos	Pecos
Colorado	Austin	Devils	Devils River
Colorado	Columbus		
International (Water) Boundary Commission			
Rio Grande	Eagle Pass	Rio Grande	Langtry
Rio Grande	El Paso	Rio Grande	Moorhead
Rio Grande	Fort Hancock	Rio Grande	Presidio
Rio Grande	Mouth of Devils River		

* Descriptions and brief histories of the stations are given in Taylor and Hoyt (1906) and IBC (1931).

for the purpose of flood warnings. This program was transferred to the Weather Bureau upon its creation in 1891 (Moore, 1896), and hydrology and flood prediction has remained a service to the present. These stations have either been integrated into the USGS network or discontinued. In 1900, there was exactly one such station in Texas, at Arthur City on the Red River (Frankenfield, 1900). The Corps of Engineers has also been active in stream gauging for many years, often in support of specific Corps projects. In Texas, this activity has been manifested mainly in funding USGS stations.

2.2.2 Proxy measurements and accuracy

The *direct* measurement of streamflow can only be effected by two methods (Jones, 2002): (1) trapping the entire volume of flow at a site on the stream for some (presumably short) interval of time, (2) directly measuring the integral

$$Q(t) = \iint u \, dA$$

in which $u(t)$ denotes the longitudinal component of current velocity and $A(t)$ the cross section of the stream channel. Except for laboratory channels or extremely small conveyances, (1) is practically infeasible. Therefore, for natural streams and rivers, excepting small channels, streamflow can be measured directly only by method (2), which has been most commonly carried out by a series of discrete measurements of longitudinal current at a network of points over the cross section, each representing a small subarea of the section, known as the velocity-area method. Otherwise, as noted in 2.2.1 above, streamflow must be estimated by some sort of proxy measurement, stage being the most common. (Much less common, and ill-suited for routine continuous monitoring, are inferences from some sort of dynamic water or mass budget.)

There are several sources of error in the field determination of discharge (Pelletier, 1988; Herschy, 2002):

- (i) variance in measurement of current speed due to meter characteristics, and orientation in the flow
- (ii) imprecision in positioning of cross section sampling points
- (iii) excessive spacing between cross section stations, i.e. too few verticals or too few sampling depths
- (iv) inadequate sampling time for current measurement (given the intensity of turbulence and frequency of pulsations in the current)
- (v) excessive elapsed time for sampling cross section (during which mean flow or transverse circulations change)

A series of these measurements is labor-intensive, and therefore cannot be performed as a routine operation. Instead, the proxy measurement of water-surface elevation is used, and occasional streamflow measurements by the velocity-area method are fitted to a stage-versus-discharge (or rating) relation for routine monitoring, thereby incurring additional sources of error:

- (vi) inaccuracy of stage measurement
- (vii) hysteresis and other pathological variations of stage with flow

(viii) scatter in the fit of stage and flow measurements to the rating curve

(ix) errors in rating-curve selection, and in reading strip-chart records, field notes, and the rating curve

Of the sources of error listed above (i) – (v), which represent the error in a single measurement of discharge, are more or less independent and operate additively, so that the total variance is given by:

$$s^2 = \sum_{k-1}^{K} s_k^2 \qquad (2)$$

where s_k denotes the standard error* of each of K component sources of error, see Whitaker and Robinson (1929), Scarborough (1966) or Topping (1962).

The relative standard error† in a streamflow measurement is frequently estimated to be on the order of 5% or less (e.g., Grover and Harrington, 1943; Pelletier, 1988; Sauer and Meyer, 1992; Herschy, 2002). Since the early development of the rotary current meter, determination of its accuracy, (i) above, has been a central issue (e.g., Henry, 1872, Murphy, 1902, Murphy, 1904). Pelletier (1988, 1990) reports a comprehensive literature review that surveys sources of error (i) – (v), above. While each specific data-collection practice requires a separate estimation of its standard error, which in turn depends upon procedures, equipment and conditions, a rough estimate can be made of the combined error using (2) above and the citations provided by

* In this report, statistical dispersion (i.e., error) is indicated by the standard error, the usual convention in physics. This is in contrast to reporting error as two standard deviations, i.e., the 95% confidence band, about the predicted or expected value. While this is a common informal practice in hydrology — though not universal (see, e.g., Sauer and Meyer, 1992) — and is recommended by the International Organization for Standardization, this writer stubbornly refuses to cave.

† The convention of quantifying discharge uncertainty with the relative standard error is also a recommendation of the International Organization for Standardization (see note, above). Though this is a common practice, and is employed in this report, one must be cautious in that the use of relative standard error assumes the error to be proportional to the magnitude of discharge. In natural streams, the standard error for some of the components listed above is independent of the measured value (i.e., homoscedastic), so the error will be underestimated at small values of the flow if an estimate of the relative standard error is applied to the value of flow. Alternatively, the relative standard error will be overestimated if it is inferred from errors of measurement at these small flows.

Pelletier (1988). This proves to be a relative standard error of about 5%, which accords with the estimate of Sauer and Meyer (1992), who opine that “most measurements [of USGS] probably will fall in the range of 3 – 6%.” However, this estimate is based upon the assumption that the observer follows standard protocols, using well-maintained and calibrated instruments, and works under ideal conditions. As noted often over the years (e.g., Murphy, 1902; Corbett, 1943; Wood, 1945; Pelletier, 1988), this is not a generally applicable assumption. The departure of field operations from ideal conditions for measurement is an especially confounding factor. Execution of flow measurements is greatly hampered by ambient conditions, notably agitation of the stream itself, particularly at higher flows. Variability of streamflow ranges from large-scale turbulence, including lateral and vertical circulations, to multiple factors influencing the slope of the water surface, all of which undermine the repeatability and representativeness of point measurements of velocity.

Sources of error (vi) – (viii), which contribute to uncertainty attending the rating curve method, have received much less attention in the literature. McMillan et al. (2010) demonstrate that uncertainty in the rating relation alone can make a considerable contribution to the error, as much as a few *tens* of percent. Clearly, any alteration in the channel geometry can modify the rating curve, and the effect of sediment scour and deposition is a widely cited factor in changes over time in the rating relation (e.g., McMillan, 2010, Tomkins, 2012). However, other factors can influence the apparent rating curve, not the least of which is the relative rarity of higher flows, which have a greater degree of leverage in estimating a rating curve. (This may also contribute to the notion that the standard error increases with the magnitude of discharge.) At the low end of the flow spectrum, in a study of rating-curve uncertainty in Australian rivers, Tomkins (2012) found that the standard error increased, particularly at low flows, due to changes in the rating relation.

Over the years, USGS has been digitizing streamflow data as well as supporting information, including field surveys, for dissemination via its National Water Information System. This resource is of inestimable value, providing not only streamflow data *per se*, but observations of channel conditions, channel cross sections, equipment descriptions and, of course, field

Table 4
Uncertainty as relative standard error (RSE) for discharge measurements at selected Texas stream gauges

River	Gauge	USGS ID	period†	"good"* or better (%)	"fair"* or worse (%)	number of msmts	RSE (%)
Sulphur	Talco	07343200	1987 - 2011	67	33	184	50
Big Cypress	Pittsburg	07344500	1987 - 2011	62	38	21	37
Little Cypress	Jefferson	07346070	1991 - 2011	54	46	134	72
Trinity	Oakwood	08065000	1991 - 2011	12	88	126	88
Brazos	Seymour	08082500	1987 - 2011	65	35	215	188
San Bernard	Boling	08117500	1995 - 2011	25	75	127	18
Guadalupe	Victoria	08176500	1987 - 2011	50	50	163	28
San Antonio	Falls City	08183500	1987 - 2011	33	67	168	14
San Antonio	Goliad	08188500	1991 - 2011	41	59	141	10
Nueces	Cotulla	08194000	1990 - 2011	41	59	98	71
Nueces	Three Rivers	08210000	1991 - 2011	52	48	152	14

* USGS data quality categories (accuracy relative to actual flow, %)

Excellent	< 2
Good	2 - 5
Fair	5 - 8
Poor	> 8

† Period of data is the period of years for which both daily-mean stage and discharge are reported by USGS.

measurements of stream discharge. USGS assigns a judgment of the quality of the field measurement of discharge, ranging from “poor” to “excellent” presumably based on conditions attending the field work. These are summarized for selected gauges in Table 4. In this study, we used the daily records of both stage (i.e., gauge height) and flow, as well as the field measurements of discharge upon which the rating curve is based. For Texas gauges, the record of daily stage is now available back to the early 1990’s or late 1980’s, depending upon the gauge.

Examples of field measurements of discharge and corresponding rating relations are shown for the Guadalupe at Victoria in Figure 7, and, for only the lower flows, in Figure 8. (The two extremely high flows of October 1998 — far above 100,000 cfs — are excluded from these plots

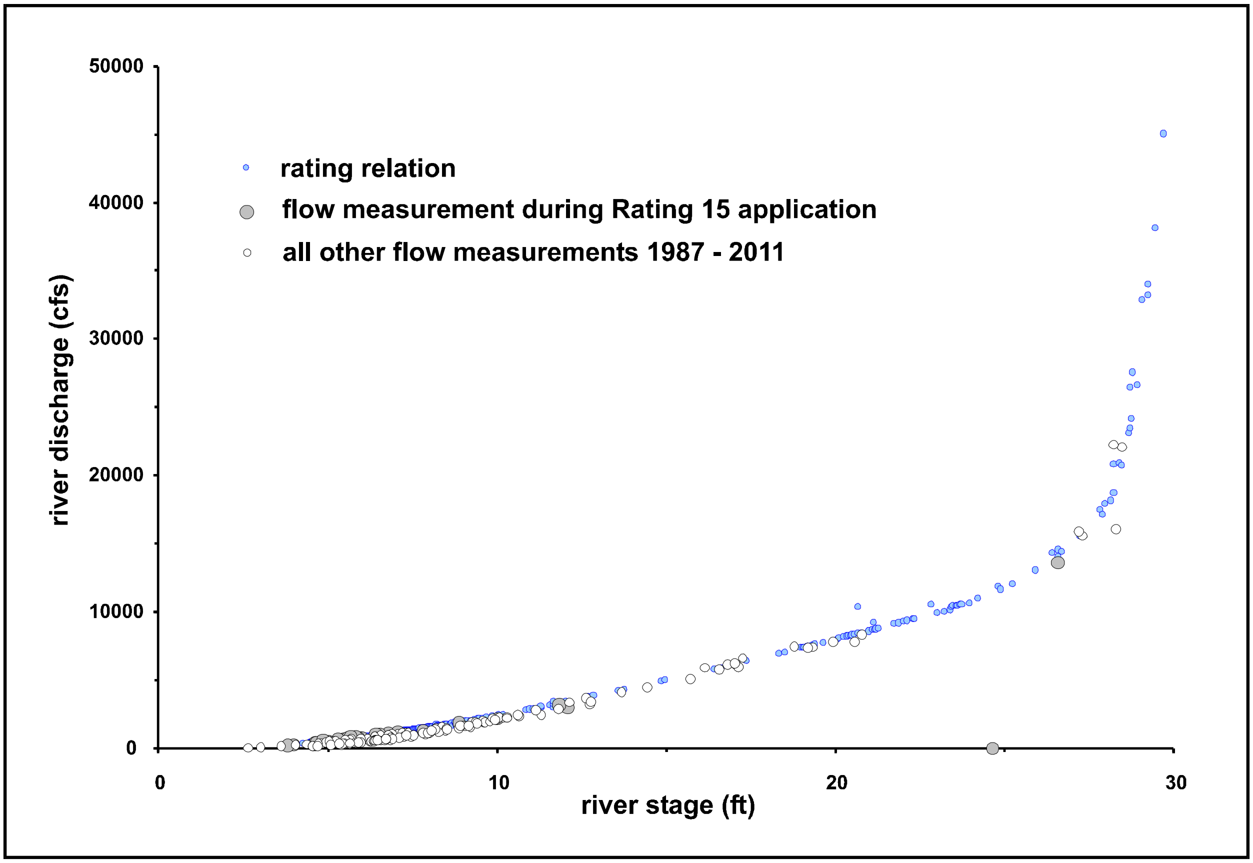

Figure 7 - Rating relation 15 (Aug 87 – Dec 91) and field measurements of discharge, Guadalupe at Victoria

to limit the axis range.) Points on the rating curve are the reported daily mean values of stage and flow (since the flow values are determined from the relation using the monitored value of stage), shown as the small blue data points on the figure. Discharge measurements during the period when the indicated rating relation prevailed are shown as large grey data points. The white data points are all other measurements in the post-1987 period (when both stage and flow data are available as digital records), when other rating relations applied. The rather abrupt excursion in flow versus stage at about 28 ft is characteristic of this gauge, because a secondary stream channel, capable of carrying an appreciable flow, becomes active at this stage.

Preparatory to the calculations of Chapter 4, an assessment of the probable uncertainty of streamflow was made at a selection of Texas gauges, based upon information available on the

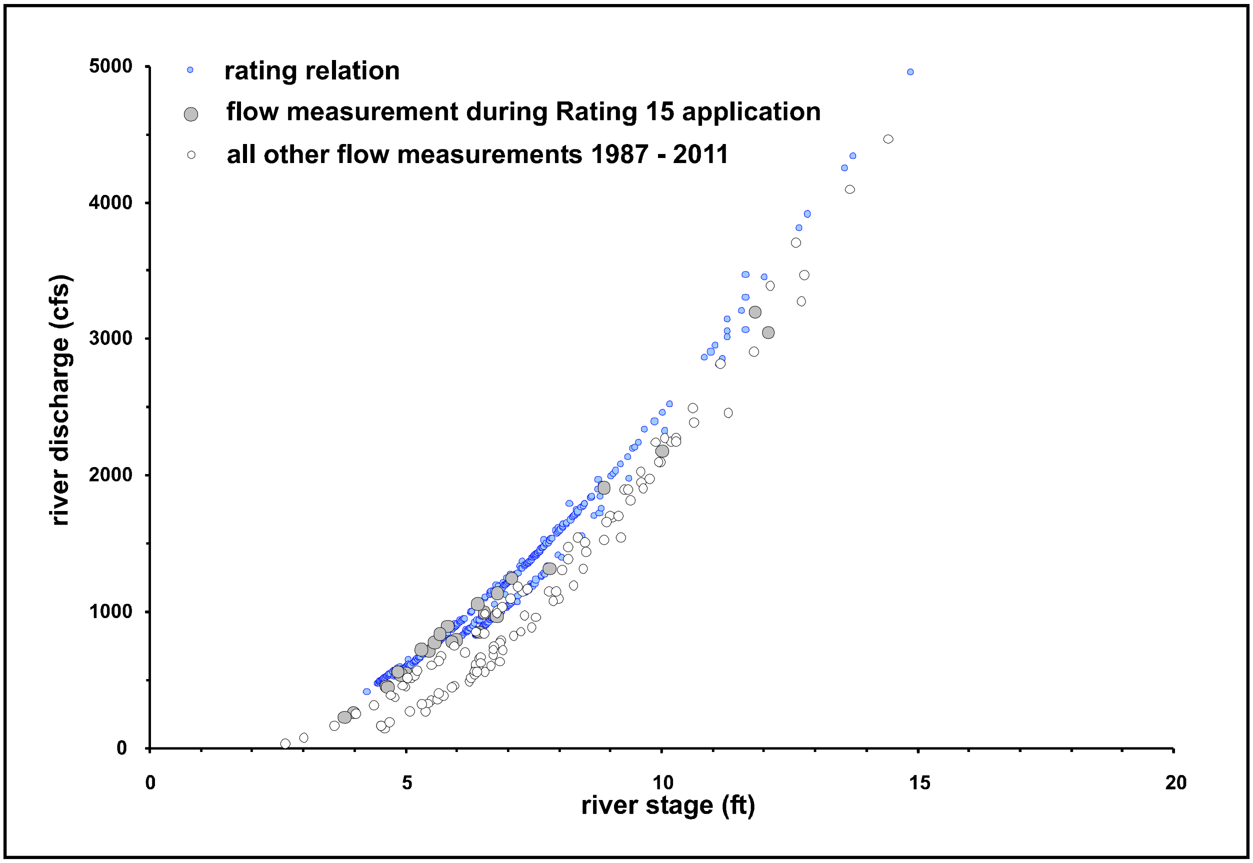

Figure 8 - Same as Figure 7 except limited to flows below 5000 cfs

USGS website. The procedure consisted of these steps: (1) segregate the record of daily stage and flow according to the prevailing ratings; (2) for each rating, create a rating table by ordering these data by stage; (3) for each rating, tabulate the field observations of stage and discharge Q during the period in which the rating applied; (4) using the field-observed stage, determine the corresponding value of flow Q_r from the rating table; (5) compute $\delta = (Q - Q_r)/Q_r$; (6) estimate the relative standard error by the mean square δ, i.e., $\sqrt{\overline{\delta^2}}$, over all the values of measured Q and over all rating relations (whose effects are normalized by Q_r).

The results are summarized in Table 4. In general, the relative standard errors are considerably larger than the above-cited sources. By the nature of the procedure used here, there is no means of separating the standard error into its component sources; rather the resulting standard errors represent the combined effects of all of the above sources of error. These field measurements are made by experienced personnel following USGS-prescribed protocols, using well-crafted, well-maintained equipment. In its appraisals, USGS rarely assigned an “excellent” rating to these field measurements, and, indeed, the high proportion of “fair” or worse indicates frequent concern with the field results. At the least, adherence to protocols and proper equipment in the field evidently is not sufficient to achieve the accuracy attainable under idealized controlled conditions, and therefore the additional sources of error enumerated above are substantial in their effect on field data. No reader familiar with performing these kinds of measurements in a real Texas stream will be surprised by this conclusion.

It is apparent that the overall error in the reported streamflow measurements is often greater than the nominal 5-10% usually cited, particularly at the lower flow magnitudes. Development of a better estimate of this error from available data would exceed the scope of this brief review. This error is worthy of much additional study, notably in its influence on water budgets involving the difference in flows between two gauging stations, such as gain/loss studies on specific river reaches, and in the interpretation of uncertainties of environmental flows of low magnitude.

3. EXAMPLES AND EXPERIENCE WITH HYDROCLIMATOLOGICAL INDICES

The three variables, rainfall, streamflow, and reservoir contents, are used to organize the literature survey, which is presented in this chapter. There are other variables that measure some aspect of water substance in the environment, such as soil moisture, water table elevation, evaporation, and transpiration, which are addressed as appropriate in the following review, but which either have limited geographical utility, inadequate instrumental record, or fail to exhibit some of the desiderata for an indicator listed in Section 1.1.1.

Hydrometeorological parameters such as these, when used as the basis for an indicator, are typically expressed as an anomaly time series, that is, a time series of departures from a *reference* condition. Definition of the reference condition becomes part of the formulation of the indicator, depicting some measure of "normalcy." The simplest such reference is a long-term mean, but the reference condition can be more complex, for example the long-term mean calendar variation. Sometimes, the anomaly is standardized, that is, the reference condition is used to define a mean and standard deviation, and the departures from that mean are divided by the standard deviation. The resulting index is dimensionless and typically ranges ± 3 - 4 about 0.

The period of record used to quantify the reference condition is referred to in this report as the *baseline* period. Selection of a baseline period is potentially crucial to constructing the index, and often reflects the personal philosophy of the analyst. Is climate fixed or does it change? If the former, then a baseline extending over the available data record would capture the full range of variation and better establish the statistics of that variation. If the latter, then the baseline period becomes a compromise between a long enough period of record to adequately define the reference conditions, but short enough to avoid errors due to underlying climatological trends.*

Practical application of an index often focuses on segments of the time series in which the index

* This issue is addressed with rigor and depth in Koutsoyiannis and Montanari (2007).

is above or below its reference condition(s). In this case, it is not the individual value of the index that is of concern so much as its systematic persistence of surfeit or deficit. Various terminologies have been employed historically for these periods in a moisture index time series, including "wet spell," "dry spell," "pluvial" and "drought." Identification of such periods clearly requires a definition of when they begin and end, which in turn necessitates threshold or trigger values of the index (or indices) for each type of condition, perhaps supplemented by other identification rules, as will be seen. Additional thresholds can be devised to communicate the relative extent of the departure from the reference.

We distinguish between the intensity of a drought or pluvial, and its severity. The former is the average departure from the reference during a pluvial or drought event. The latter is the product of this average and the duration (time of end minus time of beginning) of the event.

There are other aspects of the analysis of an index time series that will be addressed in the context of the following reviews. However, one that is ubiquitous and of particular importance in hydroclimatology is the use of a *cumulative*, that is, a running sum of the values of the series, whence its alternative name *run sum* (Beran and Rodier, 1985). Though the mathematical device has a wide use, in hydrology it has gained the name "mass curve."* The cumulative of an anomaly time series, a "residual mass curve," is especially utilitarian for its diagnostic value. This technique has a long history in climatology, being first described in detail by Marvin (1923). (Occasionally, the "mass curve" terminology is applied to the cumulative of precipitation, usually by hydrologists, e.g. Myers et al., 1969.)

* While Rippl did not use the term in his classic 1883 paper, the term may have arisen in the practice of engineering earthworks in the nineteenth century.

3.1 Precipitation-based indices

Because of its simplicity of measurement and consequent availability of data, precipitation, along with temperature, has been employed historically as an indicator of climate. For climatic classification purposes, rainfall is typically reduced to monthly and annual means, then averaged over a lengthy period of record to screen out interannual variations. The spatial distribution of these averages may yield insight into the variation of climate and the underlying physical controls. For example, the Köppen system (Köppen and Geiger, 1930), which is the foundation for modern climate classification, is based entirely upon patterns of precipitation and temperature. Helmut and Alkhalaf (1995) demonstrated that there is a direct relation between the Köppen-Geiger climate classes and the monthly components of the surface energy balance.

3.1.1 Precipitation time series and simple indicators

The simplest hydrometeorological indicator is precipitation itself, conventionally accumulated over some sampling interval Δt. As a time series, this is:

$$P_i \equiv \int_{(i-1)\Delta t}^{i\Delta t} p(t)dt \tag{3}$$

in which $p(t)$ is the instantaneous time function of precipitation rate (depth of water per unit time). The interval Δt may be the observation time interval, e.g., time between readings of a rain gauge — one day in the cooperative observer network of NCDC — or may be a further arithmetic accumulation, for which month and year are conventional periods. The use of monthly and longer averaging periods is not arbitrary, especially in Texas. At short time periods, on the order of a day, there is a marked asymmetry between extreme high- and low-precipitation events. The high-precipitation events are typically brief and intense, engendering runoff that may be sustained for several more days or perhaps weeks, depending upon the magnitude of the event and the condition of the watershed. In contrast, the low-, or, more precisely, no-precipitation events are much more numerous and clustered in time, with associated low streamflows that are highly autocorrelated. The intense, storm-driven rainfall events of course frequently entail flooding, and their measurement and analysis represent a considerable segment of the hydrological literature. In many hydroclimatological studies, including this report, the detailed analysis of individual storm events is typically avoided in favor of an analysis of averaged conditions, for which a month or longer is a suitable time interval, extended periods of which are described as "droughts" or "pluvials." It is understood that in Texas, especially, a pluvial will be made up of individual storm events and their immediate aftermath, and a drought may include occasional storm events, but the analysis addresses longer period variation for which such detail may be unimportant and therefore suppressed. Most of the literature has focused on the deficit periods — drought — rather than the pluvials, though the latter is arguably as important as the former in characterizing a region's hydroclimatology (e.g., Kangas and Brown, 2007).

As an indicator, equation (3) is site-specific and, lacking a reference, provides no temporal information other than what can be gleaned from inspection of the time series P_i itself. These facts notwithstanding, precipitation has been historically used as an indicator of wetness or dryness. In Great Britain, for example, for many years drought has been defined as a period of at least 15 days with no daily rainfall exceeding 0.25 mm (0.01 ins) rainfall, which never fails to provoke laughter in a Texas hydrology class. In the Pacific Northwest, Munger (1916) defined a drought to be a period with no daily rainfall exceeding 0.05 inches (1.2 mm). These examples illustrate three aspects of a climate indicator: (1) both the duration and magnitude of a precipitation anomaly are important; (2) the perception of what is anomalous precipitation is very much dependent upon location and the local climate; (3) for an indicator to be useful in quantifying hydroclimate, it must be coupled with specified threshold values signifying wet or dry conditions.

As an index, the time series of precipitation can be coupled with a threshold value. This threshold serves to subdivide the precipitation series into "runs" of above-threshold and below-threshold values, examples of which are given above. The duration and statistics of these runs then become the principal parameters characterizing hydroclimate. Many studies of precipitation variability have been based on the theory of runs, which was first given rigorous development by Yevdyevich (1967) and continues to be a major approach especially in the study of drought, e.g., Llano and Penalba (2011), Lei and Duan (2011). A principal difficulty with this simple threshold approach is that short duration events that "break" the run, such as storms during a drought period, or short dry spells during a pluvial, result in many short-term events, while the watershed is known to respond to the longer-term trend in moisture conditions. This can be avoided by integrating or "pooling" the time series some way before applying the threshold. Typical strategies include a moving average and the cumulative, or "mass-curve," discussed above, in which case the simple time series becomes more complicated and the associated index moves closer to complex indices to be discussed below.

A slightly more complicated version is the precipitation time series with multiple thresholds. A prominent example is the use of cumulative frequencies of occurrence, called the precipitation-decile approach. This has been widely used in Australia (Gibbs and Maher, 1967; Beran and Rodier, 1985; Kininmonth et al. (2000); Keyantash and Dracup, 2002), where it is known as the rainfall deciles-based drought index (RDDI). The decile approach is usually based on the monthly time series whose cumulative frequency is constructed and the decile points identified. The moisture categories of the Australian Bureau of Meteorology are shown in Table 5. The definition of a drought in the Australian method is more complex. It employs the monthly time series of running 3-month totals. A drought, or "serious rainfall deficiency," is considered to begin when a monthly value of the 3-month sum (i.e., the sum of that month and the two preceding) falls below the first decile (10%) of the 3-month cumulative frequency. A severe drought, or "severe rainfall deficiency," occurs when the 3-month sum falls below 5% cumulative frequency. The drought ends when either (1) the 3-month total exceeds the seventh decile (70%), or (2) the *monthly* total exceeds the third decile (30%) of the 3-month rainfall distribution.* (The latter condition is given by Keyantash and Dracup, 2002, but not mentioned by Mpelasoka et al., 2008.) Mpelasoka et al. (2008) found that the RDDI has a significant deficiency in lacking a "memory" of previous rainfall, and found an extension of the RDDI including a rudimentary soil-water budget to be more useful for water-resource management.

* One must distinguish between *decile* as a range of frequencies and as a specific frequency. Above, and elsewhere in this report, the latter usage is observed. The *n*th decile in this convention is a value of the variable (rainfall, above) corresponding to a cumulative frequency of $n/10$ where n is an integer. Thus, the first decile is the value of the cumulative distribution at 0.10 (10%), the fifth decile is at 0.50 (50%) and the tenth decile is at 1.00 (100%). In the former convention, used, e.g., by Gibbs and Maher (1967) and Kininmonth et al. (2000), the first decile are the values with frequencies ≤ 0.10, the fourth decile are those with $0.3 < \text{frequency} \leq 0.40$, and, generally, the *n*th decile are those values with frequencies in the range $(n-1)/10 < \text{frequency} \leq n/10$.

Table 5
Moisture condition decile classes of annual and monthly precipitation used by Australian Bureau of Meteorology, from Kininmonth et al. (2000), see also Gibbs and Maher (1967), and Bordi et al. (2001)

designation	*cumulative frequency value**
very much below average	≤ 0.10
below average	(0.10, 0.30]
average	(0.30, 0.70]
above average	(0.70, 0.90]
very much above average	(0.90, 1.00]

* $(a,b]$ designates a numerical interval inclusive on the right, that is, all values x such that $a < x \leq b$.

The decile method was used by Chakravarti (1976) to study drought on the Canadian prairies. Chipanshi et al. (2006) studied climate extremes in this same region — mainly focusing on drought — as defined by the upper and lower deciles of exceedance frequencies of annual-total (the agricultural year of 12-months starting in September) and growing-season total (6-months April through August) precipitations. A drought was defined to begin when the total annual or seasonal precipitation for that year was less than the corresponding lowest decile, and to end when the precipitation exceeded the median.

Sometimes the indicator thresholds are specified in terms of a reference value $\hat{P}$, most often a statistic such as the mean or median, or some fraction thereof. This offers a modicum of compensation for temporal, station or regional differences in climate, by relating precipitation to the reference, which is regarded as “normal” (Dracup et al., 1980). Even though precipitation may vary widely in space and time, its magnitude as a fraction of “normal” and or its recurrence as an exceedance frequency are measures of departure from “normal” conditions, which are often assumed to be invariant over time and space. There is, of course, little basis for this assumption. In this context, “normal” is a local perceived condition considered characteristic, based generally on a period of record of observations in the area. There is also a formal definition of climatological “normal” presented below.

The natural and most widely used such reference is the average precipitation $\overline{P}$ over an extended period of several to many years. This characterizes the local rainfall by a single number, the mean annual. A slightly more complex reference is the long-term mean monthly $\overline{P}_j$, for $j = 1, \ldots, 12$ (1 denoting January, *et seq*.), which has the additional advantage of accounting for the seasonal variation of precipitation. Like the long term annual mean, the long-term mean monthly has been used for many years as a climatological reference.

The climatological "normal", the averaged values, both annual and monthly, over a 30-year period beginning with year 1 of a decade (e.g., 1901, 1911, etc.), is a common reference average. The use of 30 years as a baseline period can be traced back to the meeting of the International Meteorological Committee meeting in 1872, though it was not until 1935, at the Commission for Climatology of the International Meteorological Organization (the predecessor to the World Meteorological Organization, WMO) in Sopot-Danzig, that the decision was reached to establish the 1901-1930 period as the baseline for climatological studies (World Climate Programme, 1989; Commission for Climatology, 2011).* Later, the WMO specified that the normals would be re-calculated every 30 years, so that 1901-30, 1931-60, etc. are the "standard normals". In the U.S., the normals are re-calculated every decade. At this writing, the current normal is in transition to 1981-2010. As noted in the introduction to this chapter, there are philosophical issues lurking beneath the selection of a baseline period, whether changes from one baseline period to another are due to vacillations in a stable climate or to the trend of a mutable climate, and, if the former, whether the average can be improved by a longer baseline (see, e.g., Guttman, 1989; Arguez and Vose, 2011). In Texas, Figure 9, the changing 30-year normals suggest a trend of increasing precipitation and possibly a shift in seasonality.

The fraction-of-normal, $P\,/\,\overline{P}$, or percent-of-normal, perhaps the simplest precipitation-based indicator, has had extensive historical use, and continues to be employed to the present (e.g., Quiring, 2009a, 2009b). A closely related index is $P\,/\,\overline{P}\ - 1$, which is equivalent to

* This decision was over the opposition of the North American representatives.

Table 6
Drought thresholds used in India with index (4), from Shewale and Kumar (2005) and Sarkar (2011)

designation	$(\overline{P} - P)/\overline{P}$ *value**
drought	≤ 0.25
moderate drought	(0.25, 0.50]
severe drought	> 0.50

* (a,b] designates a numerical interval inclusive on the right, that is, all values x such that $a < x \leq b$.

$$\frac{P - \overline{P}}{\overline{P}} \tag{4}$$

where P can be monthly, seasonal or annual rainfall, and $\overline{P}$ the corresponding mean. This is the precipitation surfeit as a fraction of the mean. If the index is used to track a deficiency of moisture, it is often written as the moisture deficit as a fraction of normal, i.e., the negative of (4), and in this form called the rainfall deficit index (e.g., Naresh Kumar et al., 2009).

India has used this index, averaged over the monsoon season (June – September, see Chaturvedi, 1985), since 1875 to track relative moisture in its climatic divisions (Shewale and Kumar, 2005; Sarkar, 2011). The principal concern is drought, for which the thresholds are given in Table 6. Mooley (1994) proposed combining this with criteria for the area affected by drought, creating joint criteria, but the resulting index is more complicated.

With a suitably defined reference value $\hat{P}$, the precipitation anomaly

$$P_i \text{ - } \hat{P} \tag{5}$$

is a fundamental indicator for hydroclimate, see, e.g., Jones and Hulme (1996). As above, $\hat{P}$ is almost always a long-term mean, often a climatological normal, and can be either the annual

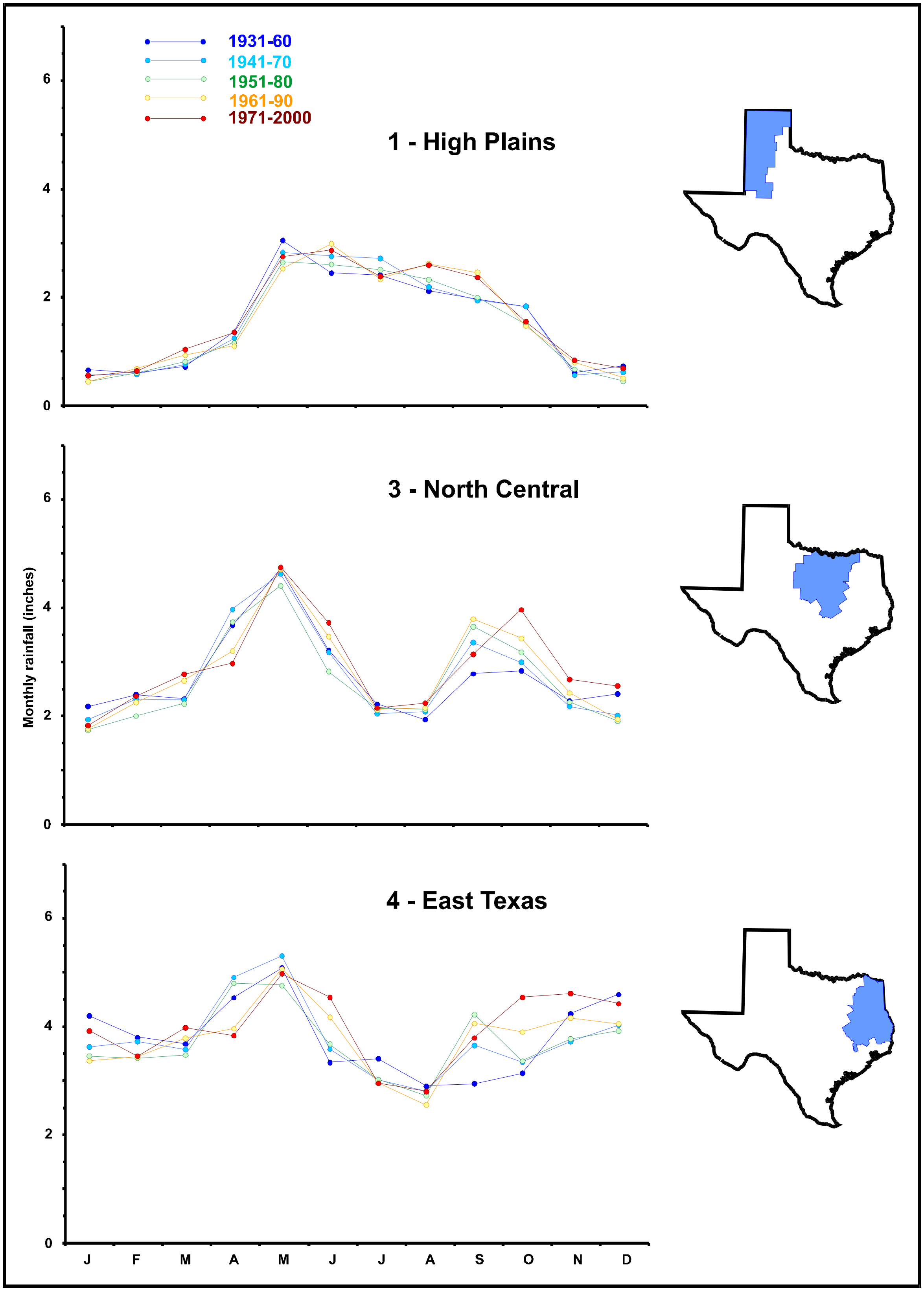

Figure 9 - Monthly precipitation in selected climatic divisions for five recent normals

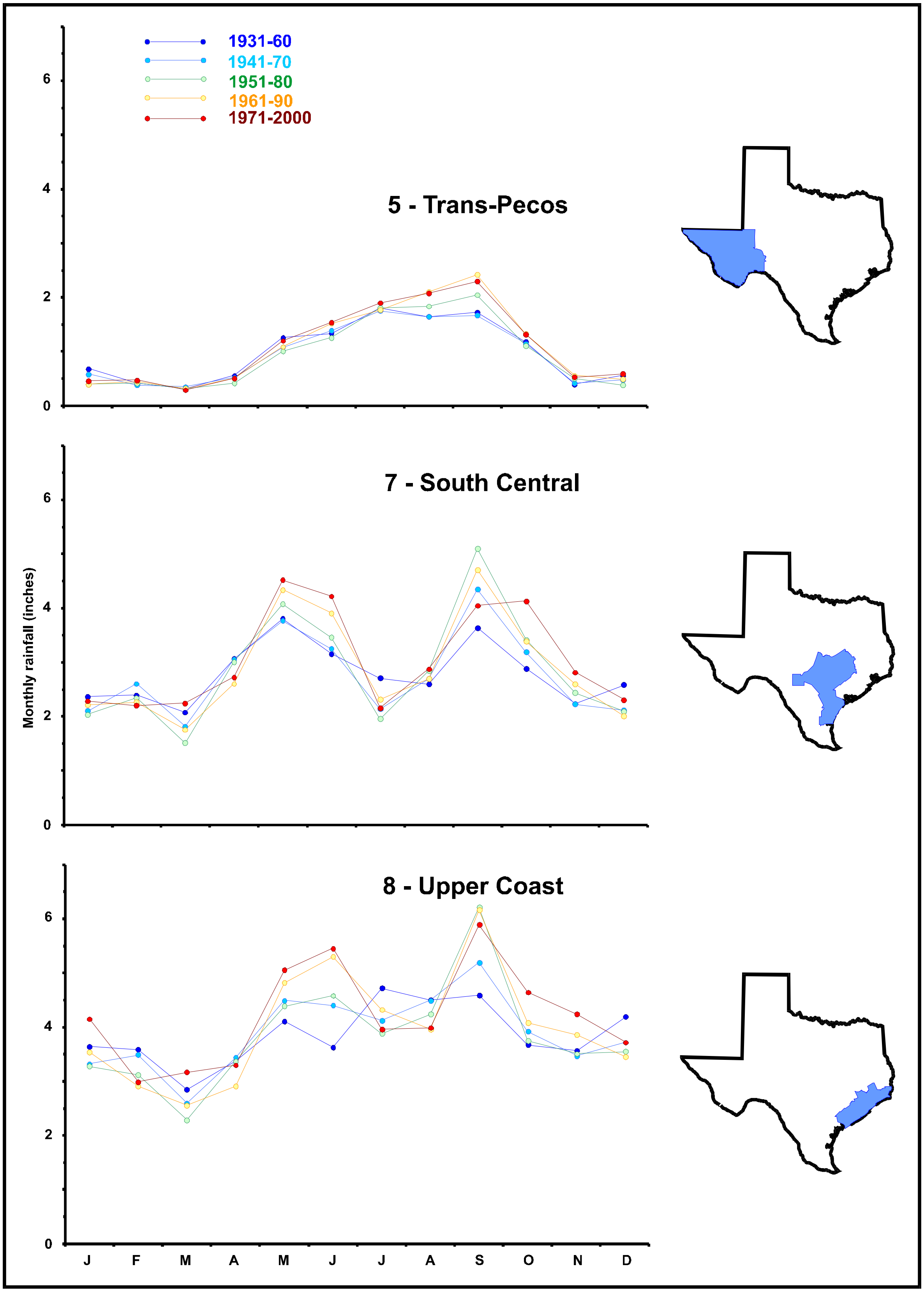

Figure 9 – (continued)

mean or monthly mean, where the month index $i \equiv j \pmod{12}$, $j = 1,\ldots, 12$. Like percent of mean, this is such a natural indicator that its first use is untraceable. For the U.S. national weather service, in its earliest climatological summaries in the postbellum nineteenth century, the Signal Service referenced the difference between monthly precipitation and the monthly mean as a measure of "excess" or "deficiency" (e.g., U. S. Signal Service, 1873).

3.1.2 Standardized precipitation index

An important variation on the cumulative of the precipitation anomaly (5) is the standardized precipitation index (SPI), which has become popular since its introduction by McKee et al. (1993, 1995, see also Edwards and McKee, 1997) as an index to drought conditions in Colorado, so much so that the NCDC now issues an updated computation of the SPI by climatic division as part of TD-9640. There are two components to the calculation of the SPI. The first carries out a calculation of the accumulated precipitation over a sliding time window of selected durations, referred to as "time scales". The second scales the cumulative precipitation relative to a long-term mean-cumulative to units of standard deviation, employing a mathematical transformation to the standard Gaussian* distribution, with zero mean and unit standard deviation.

The strategy of scaling rainfall anomaly to a statistic of dispersion has precedent in the literature. A half-century ago, van Rooy (1965) defined an analogous index for South Africa, except he used the departures $M - \overline{P}$ and $m - \overline{P}$, M and m denoting the mean of the ten highest and lowest monthly rainfalls on record, resp., as measures of dispersion rather than the standard deviation. This index is sometimes referred to as the rainfall anomaly index, or Rooy anomaly index (RAI). Kraus (1977) carried out an analysis of rainfall variation in the overland segments of the Intertropical Convergence Zone. He employed as an indicator $(P - \overline{P})/\sigma$ for each station, where P is an annual rainfall, $\overline{P}$ the average annual over the period of record, and σ the standard

* The term "normal distribution" is preferred, but in the present hydroclimatological context "normal" has a separate and important meaning. To avoid confusion — on at least this point — the statistical distribution is the Gaussian for this report.

deviation*, similar to the SPI. In a study of the Indian monsoon, Bhalme and Mooley (1980) likewise defined a moisture index as the departure of a monthly rainfall from the monthly mean divided by the standard deviation, $(P - \overline{P})/\sigma$. They further scaled this index to extreme dry periods using the same procedure as Palmer (1965, see Section 3.1.3, below) so that the index ranged generally from -4 to +4. This is referred to as the Bhalme-Mooley Index (BMI). Focusing on the monsoon period June – September, they defined a drought area index as the percentage of India's area with moisture index less than -2. Katz and Glantz (1986) and later Jones and Hulme (1996) examined some of the properties of the monthly standardized index $(P - \overline{P})/\sigma$, calling it the standardized anomaly index (SAI). They identified potential problems in averaging several stations together (as Kraus did in his study) if these stations exhibit markedly different rainfall characteristics or if there is nonhomogeneity due, say, to substantial periods of missing data. They note that these as well as other indices exhibit aberrant behavior for data sets with low precipitation values. The SPI shares these issues, as well, as will be seen. More recently, Soulé (1990) used a monthly standardized index $(P - \overline{P})/\sigma$, computed for each of the 344 divisions in the contiguous U.S. for the period 1931-85, to examine large-scale drought characteristics. Muñoz-Díaz and Rodrigo (2005) used the same standardized index to examine associations between El Niño and rainfall in Spain. Ali et al. (2009) used this index in a study of rainfall in the Sahel. They identified this index as the SPI, but it is not computed following the protocols of Section 3.1.2.1 (below).

3.1.2.1 Formulation

The SPI is determined from the time series of monthly precipitation P_i for the baseline period of record of length N months, typically but not necessarily a 30-year (360-month) normal. An integration period, or "time scale", of M months is selected. McKee et al. (1993), used M = 3, 6, 12, 24, and 48 months. NCDC provides the climatic-division monthly SPI for a range from 1 to

* In numeric application to a data set, both the average and standard deviation are those of the sample, which is an estimate of these parameters for the underlying probability distribution. This report is already over-replete with symbols, so we will not make this distinction symbolically, and therefore can reserve *s* for other uses, such as the standard error of data fitted to a regression.

24 months. This period is used as a moving window over which the SPI is computed. First a series of subsets is extracted from the monthly time series, namely

$$S_{M+(k\text{-}1)} = \left\{ P_i \,\middle|\, i = k, \quad , M+(k\text{-}1) \right\} \text{ for } k = 1,\ldots, N\text{–}M+1 \tag{6}$$

Each subset is a sequence of M values of P, the subset being indexed by the *last* value of i in the sequence. For example, if $M = 6$, then these subsets are S_6, S_7, S_8, …, S_N. For each subset S_m, the cumulative is computed:

$$S_m = \sum_{i=m-M+1}^{m} P_i \quad \text{for } m = M, M+1, \ldots, N, \tag{7}$$

where for simplicity the same symbol is used for the subset of a sequence of P_i and its cumulative. (The average of the P_i in each subset could be computed just as readily, by dividing the cumulative by M. The ultimate result is qualitatively the same.) This yields a time series that is the set of N–M+1 sequential values of the cumulative, which can be subjected to statistical analysis. It would also be possible to stratify the cumulatives by month, i.e., sort into subsets of S_m for $m \equiv j$ (mod 12), $j = 1,\ldots, 12$, if one wished to examine the statistical behavior of a particular calendar month. One should not, however, lose sight of the fact that each such month represents the cumulative rainfall over an M-month period ending in that month.

At this point, the set of M-month cumulatives could be standardized by transforming each cumulative S_m into the index $(S_m - \bar{S}_M)/\sigma_M$, for the mean $\bar{S}_M$ and standard deviation σ_M. This would transform the values of the cumulatives so that their mean is zero and their standard deviation is unity. As a Gaussian-distributed variable would have zero mean and unit standard deviation by the same transformation, it would be tempting to expect other properties of the standardized Gaussian to apply as well, notably higher moments about the mean such as skew and kurtosis. There is, unfortunately, a minor fly in the ointment, in that rainfall, and monthly rainfall, and a cumulative of monthly rainfall, are not generally Gaussian-distributed, but rather skewed positive. McKee et al. (1993) met this problem by fitting the ogive* of the set of cumulatives to that of a gamma distribution $G(S)$,

$$G(S) \;=\; \frac{1}{\beta^{\alpha}\Gamma(\alpha)}\int_0^S x^{\alpha-1}e^{-x/\beta}dx \tag{8}$$

then determining $\mathscr{N}^{-1}\{G(S)\}$ where S is the value of the M-month rainfall (whose discrete values are the time series S_m) and $\mathscr{N}$ is the ogive of the standardized Gaussian distribution. This is the SPI. Other functions besides the gamma may be used. Details of the procedure are presented in McKee et al. (1995), Edwards and McKee (1997), Guttman (1999), and Naresh Kumar et al. (2009). Several workers, e.g., Vidal et al. (2010), stratify the results by month (as noted in the preceding paragraph) and fit a distribution (typically, the gamma) by month.

We note that the ogives of both the Gaussian and gamma distributions are monotonic functions. This means that the order of the data points is preserved by the above transformation and therefore also the exceedance frequencies of individual data points (which, of course, is the underlying constraint on the transform). More information about transformation of distributions may be found in standard texts, e.g., Mood (1950), and Kendall and Stuart (1963). The SPI procedure for fitting the gamma function was taken directly from Panofsky and Brier (1958, pp. 42-43), and the transformation to the standardized Gaussian from Abramowitz and Stegun (1964, p. 1046).

Two comments are in order about the rôle of the distribution used to implicitly transform the variables S_m to a Gaussian distribution. First, the strategy of fitting a cumulative distribution function is based on the expectation that the fitted function will give a more reliable estimate of the frequencies of occurrence of the values of S_m, especially at the tails of the distribution. It can be argued philosophically that the distribution of the data themselves is the best estimate. That is, the empirical cumulative distribution should be used instead of one of the standard functions

* An ogive is a cumulative frequency distribution. Use of the term "ogive" above avoids the temptation to refer to the cumulative frequency of an accumulation of cumulatives.

Table 7
Moisture condition classes for SPI,
see McKee et al. (1993, 1995), Hayes et al. (1999)

designation	SPI *value**
extremely wet	≥ 2.0
very wet	[1.5, 2.0)
moderately wet	[1.0, 1.5)
near normal	(-1.0, 1.0)
moderately dry	(-1.5, -1.0]
severely dry	(-2.0, -1.5]
extremely dry	≤ -2.0

* $(a,b]$ designates a numerical interval inclusive on the right, that is, all values x such that $a < x \leq b$. $[a,b)$ denotes an interval inclusive on the left.

(e.g., Shukla and Wood, 2008). In this case, the SPI becomes $\mathcal{N}^{-1}\{F(S_m)\}$ where $F(S_m)$ is the cumulative frequency of S_m. The plausibility of this depends mainly upon the degree of skew in the S_m data, particularly if there are only a few data on the tails of the distribution. Second, it should be noted that the S_m data are considered to be drawn from the underlying function similar to (8), so that the set of S_m is only one realization of such sampling. A little monte-carlo experimentation with drawing 30-member sample sets from a gamma or Gaussian distribution (both readily available in EXCEL®) will convince the reader that there is considerable variability in the sample, especially in the extreme values, and, therefore, a similar variability in the parameters of the fitted function. The expectation that the fitted function provides a more reliable estimate may be optimistic. The uncertainty can, of course, be estimated by statistical methods, but this seems to have been rarely, if ever, carried out.

With respect to thresholds, McKee et al. (1993) suggested a classification of the departure of the rainfall index from the average (i.e., the sigma level). Although described as arbitrary, it has been generally adopted among users of the index. This is shown in Table 7, and diagrammed in Figure 10. This figure is, of course, simply a graph of the standardized Gaussian distribution, the data for which can be had in any statistics book. What is significant is that drought and pluvial

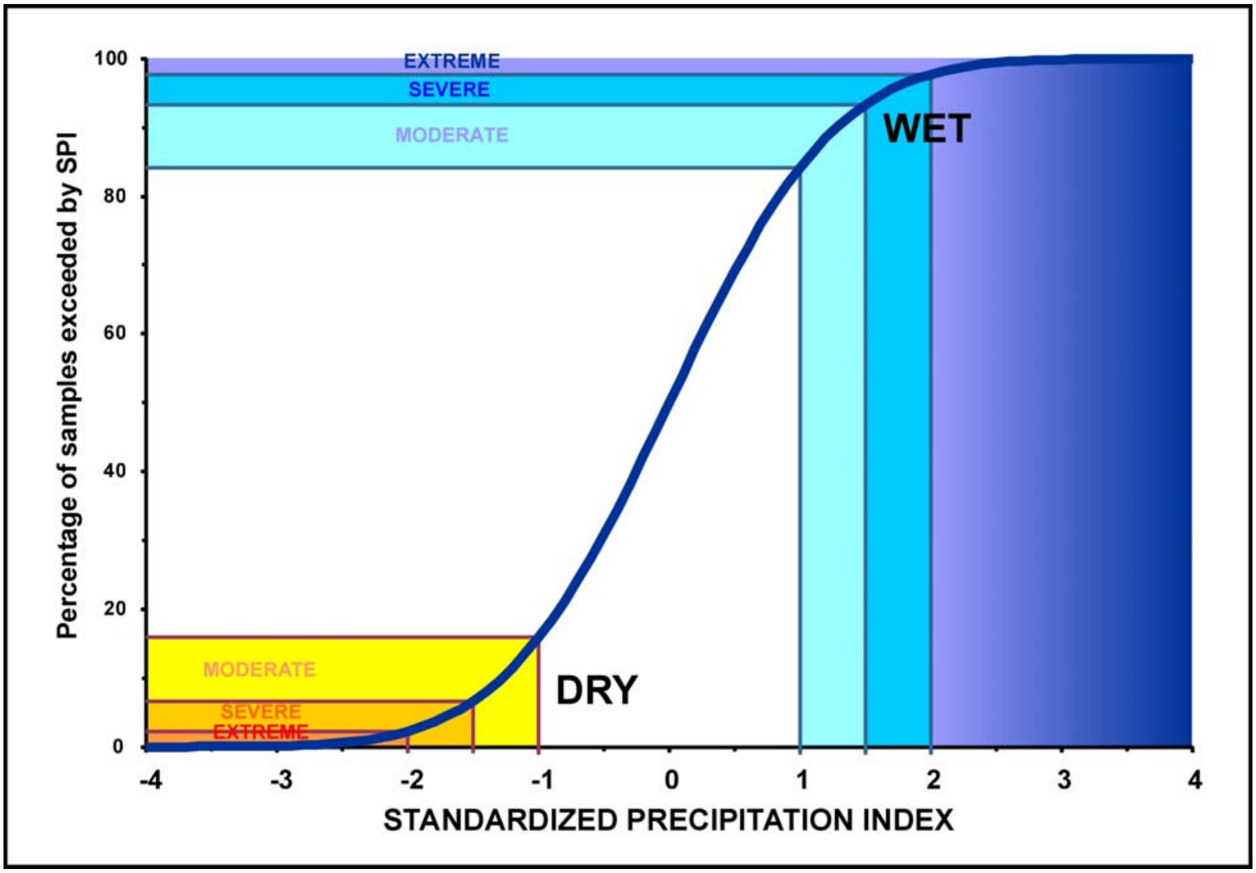

Figure 10 - Cumulative frequency distribution of SPI showing drought/pluvial-intensity categories of McKee et al. (1993)

intensities are tied to samples lying outside the 1-standard-deviation bounds of the distribution. About two-thirds of the data are thereby excluded from being classified as anomalous rainfall. (Actually, in McKee et al., 1993 and 1995, any SPI between -1 and 0 — which corresponds to the median as well as the mean — was classified as "mild drought." Guttman, 1999, assigns the term "neutral" to data between ±1, i.e., within one standard deviation of the mean. Hayes et al., 1999, Bordi et al., 2001, and Kangas and Brown, 2007, label the -1 to +1 band "near-normal.") By 1999, "dry" generally replaced "drought" in the SPI categories (e.g., Guttman, 1999; Hayes et al., 1999; Bordi et al., 2001), as observed in Table 7. It is, of course, acceptable to choose specific cumulative frequencies, rather than tidy multiples of the standard deviation, as wet/dry categories, for example 0.05 (5%) for which the SPI is -1.65, 0.20 for which the SPI is -0.84, and 0.70 for which the SPI is +0.52, etc., e.g., Andreadis et al. (2005) and Vidal et al. (2010), who employ the second decile (20%), i.e. SPI = -0.84, not SPI = -1.00, as the criterion for drought conditions.

While the time-scale governs the period of accumulation of the data, and negative values of SPI quantify the intensity of a dry spell, additional rules are needed to define a drought *per se*. McKee et al. (1993) define a drought as "a period in which the SPI is continuously negative and the SPI reaches a value of -1.0 or less." The duration of the drought extends from the first negative value of SPI to the first positive value thereafter. High variability of the SPI, characteristic of short time scales, is more likely to create short-term droughts, while prolonged droughts will be exhibited by SPI's based on longer time scales. Indeed, McKee et al. (1993) show that as the time scale (i.e., accumulation window) increases, there are fewer droughts but of longer durations. The "magnitude" of a drought ("severity" in the terminology of this report) is defined to be

$$D = \sum_{m=m_o}^{m_1} S_m \tag{9}$$

where m_o is the first month with SPI < 0 and m_1 is the first month afterward with SPI > 0. (There is at least one intervening value with SPI ≤ -1, otherwise this is not a drought period.) The units of D are months.

3.1.2.2 Properties and application

The calculation of SPI above yields a continuously valued index. The range and behavior of the SPI are greatly dependent upon the selection of accumulation time. This in effect is equivalent to subjecting the precipitation time series to a moving average of prescribed duration. Longer window durations will have the effects of low-pass filtering the input time series, which will also limit the range of variation of the SPI, and lagging in time. This ability is considered by many to be the primary advantage of the SPI: it allows the selection of a time scale for investigation to be entirely in the hands of the user (see, e.g., Vicente-Serrano et al., 2011), and other precipitation and hydrological indices have been modeled on this feature of the SPI, as will be seen. In the

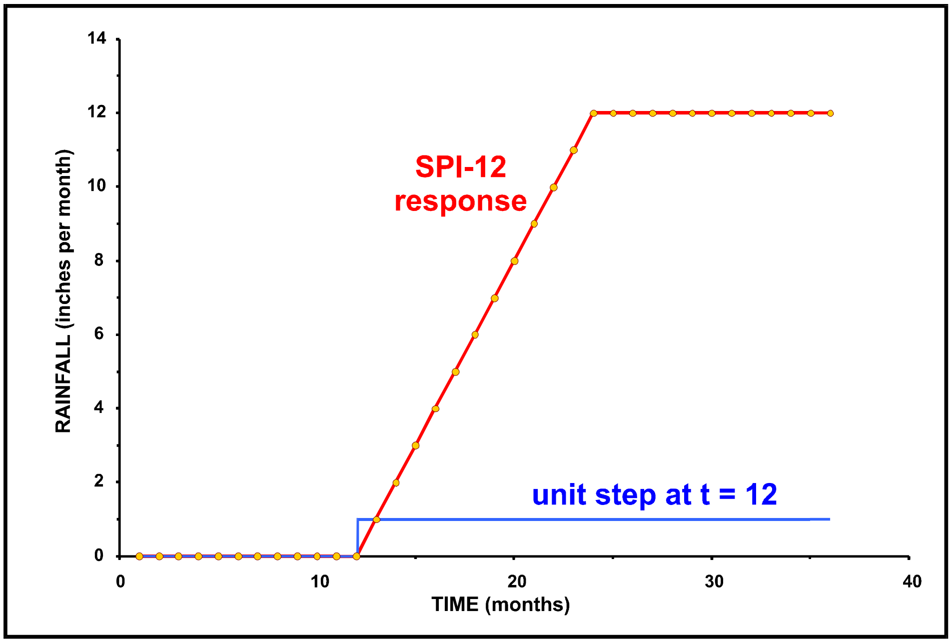

Figure 11 - Unit-step response of SPI-12 cumulative equation (7)

present context, we consider this aspect of the SPI to be essentially the choice of a time filter for pre-processing the hydroclimatological data used in the index. By re-indexing (7)

$$S_m = S_{M+n} = \sum_{i=n+1}^{M+n} P_i = \sum_{i=1}^{n} h_{M+n-k} P_i + \sum_{i=n+1}^{M+n} h_{M+n-k} P_i = \sum_{i=1}^{M+n} h_{M+n-k} P_i = \sum_{i=1}^{m} h_{m-i} P_i$$

for $n = 0, 1, \ldots, N\text{-}M$, we see the kernel h_i is given by

$$h_i = \begin{cases} 1 & i = 0, 1, 2, \ldots, M\text{-}1 \\ 0 & i = M, M+1, \ldots, N \end{cases} \tag{10}$$

The response of the cumulative S_m for M=12 to a unit step at t = 12 is shown in Figure 11. (Were this an average rather than a sum, the response would acquire the value 1.) The second component of the SPI determination, i.e., fitting a skewed distribution to the ogive and transforming to a standardized Gaussian, cannot be carried out meaningfully for a step function in rainfall. Since the kernel does not decay but drops immediately to zero after M months, it is a

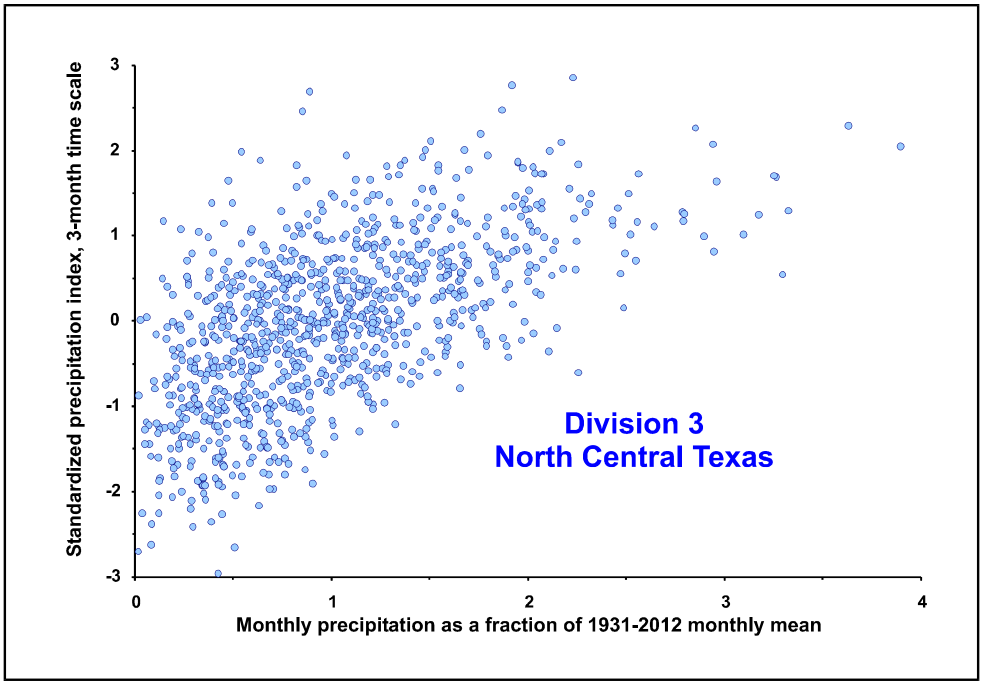

Figure 12 - SPI versus the ratio of monthly rainfall to monthly-mean rainfall

stretch to estimate a settling time and time constant, but these prove to be 11.5 months and 4 months, respectively, for whatever they might be worth.

The SPI has enjoyed wide acceptance since its introduction by McKee et al. (1993). This is one of the front-running indices recommended for use in Texas by Quiring et al. (2007). As noted above, in addition to the Palmer index, NCDC now issues updated computations of the SPI by climatic division as part of TD-9640, for time scales of 1, 2, 3, 6, 9, 12 and 24 months. Of course, for a time scale of 1-month, the set of cumulatives (4) collapses to the time series of monthly rainfalls *per se*. It is sometimes reported in the literature that at the short time scales, the SPI is similar to the percent-of-normal depiction of wetness/dryness. Such a statement should be taken *cum grano salis*, see Figure 12.

Some of the properties of the SPI emerge from examination of Figures 13 and 14, displaying the SPI's for two climatic divisions in Texas, the North Central (3) and South Central (7), see Fig. 6. The times series of SPI for time scales (i.e., accumulation windows) are plotted for 3-, 6-, 9-, 12-

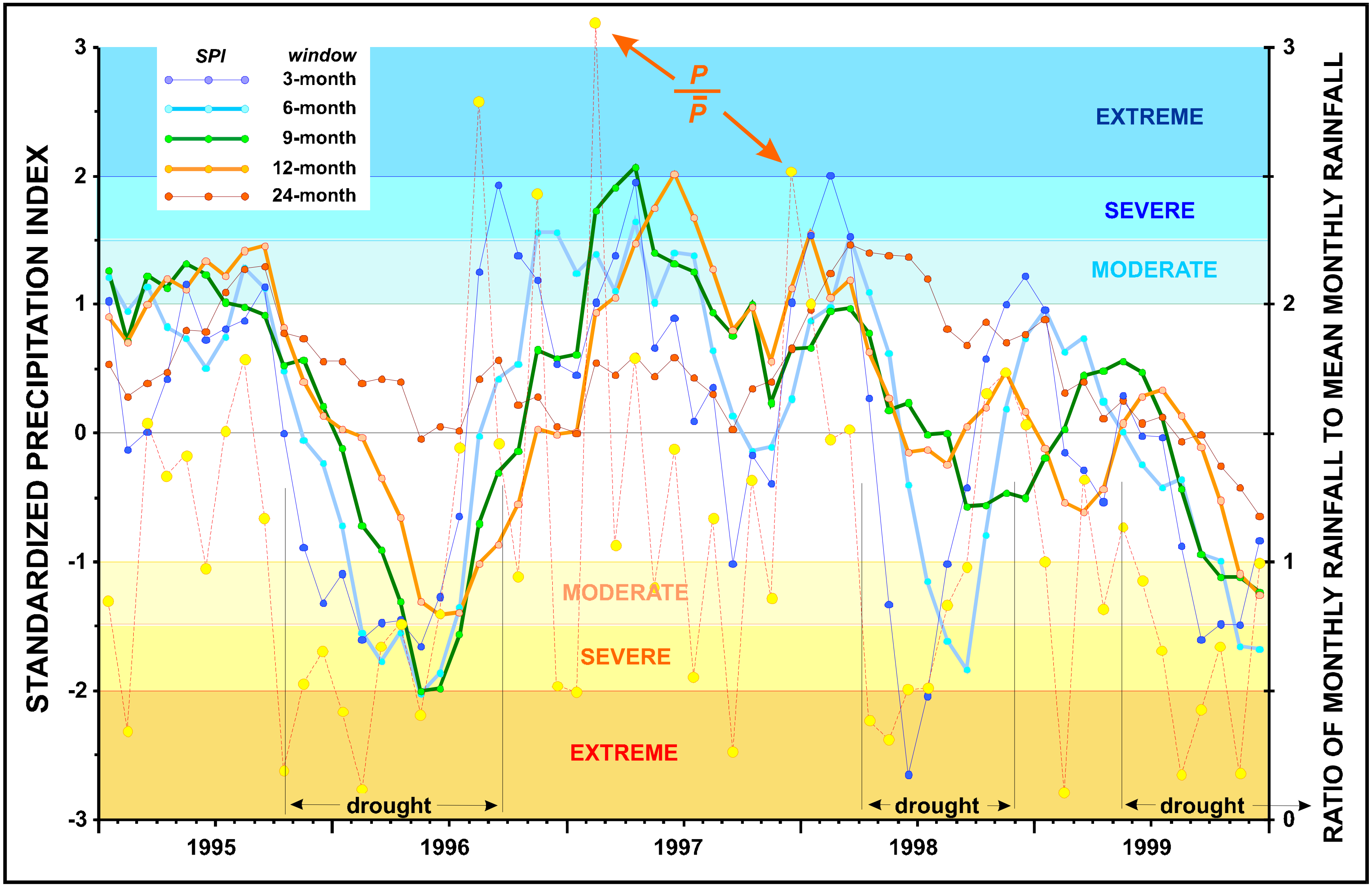

Figure 13 - SPI's for various time scales with rainfall/monthly mean rainfall, North Central Texas (Division 3), 1995-2000. Drought categories from Hayes et al. (1999).

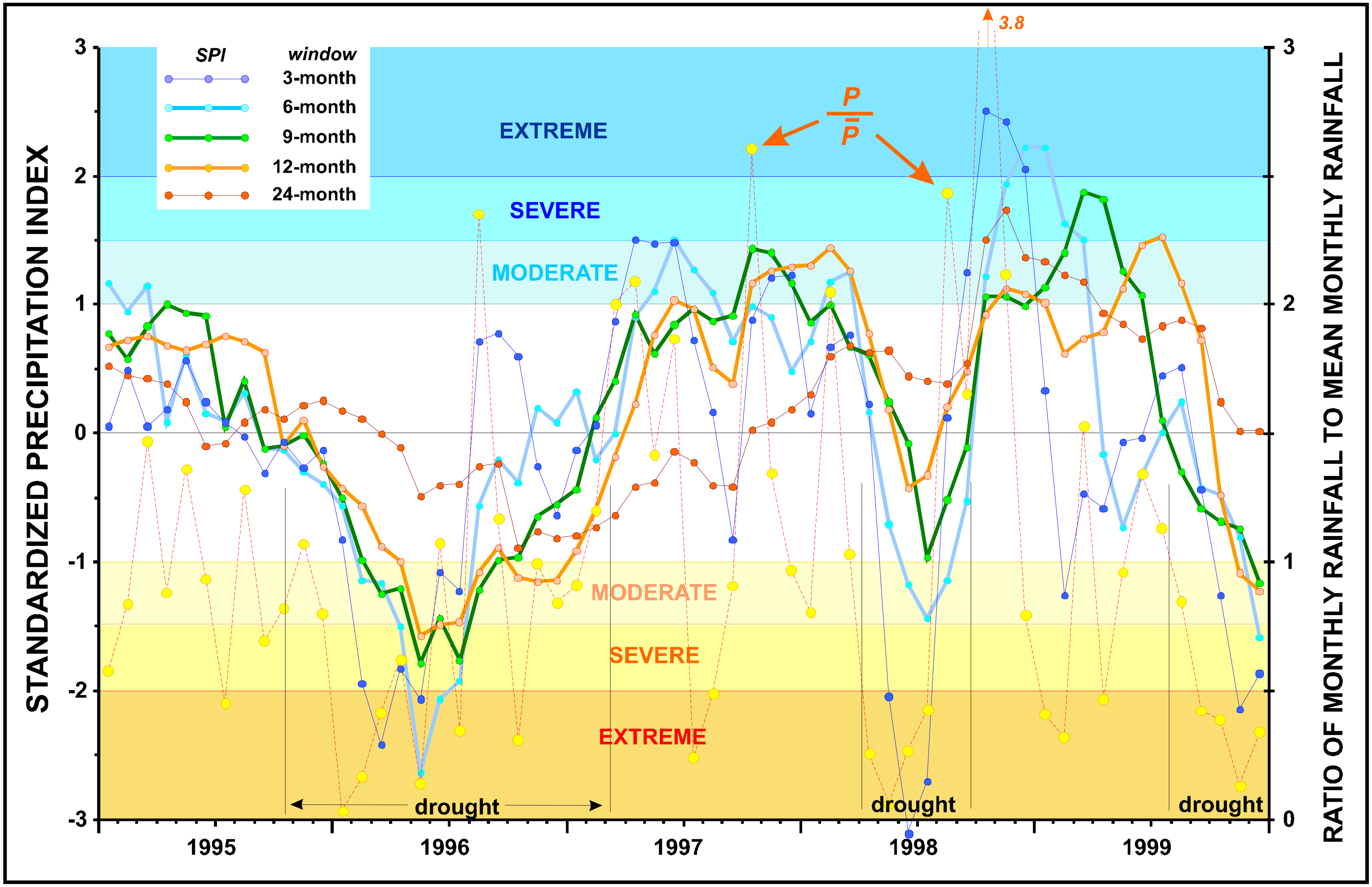

Figure 14 - SPI's for various time scales with rainfall/monthly mean rainfall, South Central Texas (Division 7), 1995-2000 Drought categories from Hayes et al. (1999).

and 24-month periods. In addition, monthly precipitation is shown as a ratio to the monthly mean for 1931-2012, i.e., the average of that month's rainfall from each year in the period 1931-2012. Three drought periods occur in this five-year period, roughly demarcated by the vertical lines. The rainfall as a fraction of the mean and the SPI-3 share the problem of fluctuating too wildly in immediate response to rainfall events. In South Texas, for example, during the first drought period both of these indices shift upward out of the drought category in July 1996, only to fall back into drought by October (Fig. 13). The SPI-24 has the opposite problem of too much smoothing and lag, barely reaching the threshold of "moderate" only after the worst of the drought had moderated. In North Texas (Fig. 14), the SPI-24 does not even fall into the drought category. Yet this drought was, at the time, of record intensity: the period October – May was the second driest such period on record, and January - May 1996 the driest on record (Halpert and Bell, 1997).

The SPI-6, SPI-9 and SPI-12 are fairly consistent in tracking the '96 drought in both divisions. They are also consistent in the duration of the '98 drought in South Texas but widely variant in its intensity, with SP-9 descending only barely to "moderate" and SP-6 only barely to "severe" (while the SP-3 drops below -3, well into the "extreme" range). In North Texas, there is no consistency among these three indices, two not even qualifying as "dry," yet the '98 drought brought worse monetary damages than the '96 drought (e.g., Smith, 1998). In many respects, these performances represent an extreme test, a short but intense dry period. The performance of these indices for longer dry periods will be examined in Chapter 4. The increasing lag in longer-period windows is not apparent if the analysis considers only interannual variation, for example, tracking the different indices only at a specific month in the year (e.g., Rouault and Richard, 2003). But this also reduces the available data to fit the distribution by a factor of 12.

One of the first published examples of the practical utility of the SPI was a study of the 1996 drought by Hayes et al. (1999). This was a relatively short drought, of about one year duration, but considered unusually severe, afflicting mainly the South Plains and southern Texas. Hayes et al. give particular attention to the situation in Texas, using the SPI's for the ten state divisions reported by the NCDC (Fig. 6). Based mainly on maps of the 3-month and 6-month SPI, Hayes et al. argue that the SPI identified the onset of drought early and the subsequent termination of drought almost at once. The "onset" and "termination" of the drought, however, seem to be inferred from the initiation and end of published accounts of the drought's effects in the news media (crop damage, wildfires, etc.). These sorts of reports may be useful for establishing the "triggers" of a drought index, by ensuring that "severe" or "exceptional" (or whatever terminology is used) drought categories correspond to "severe" or "exceptional" impacts, but are are too imprecise for determining when a drought actually began or ended.

The index has been applied in Spain (Vicente-Serrano and López-Moreno, 2005; Mestre, 2011), Italy (Bordi et al., 2001; Mendicino and Versace, 2007)), Germany (Cloppet, 2011), and the entirety of Europe (Lloyd-Hughes and Saunders, 2002). The index has also been employed in India (Patel et al., 2007; Naresh Kumar et al., 2009; Sarkar, 2011), South Africa (Rouault and Richard, 2003), Brazil (Sentelhas, 2011), Iran (Tabrizi et al., 2010) and Turkey (Türkeş and Tatlı, 2009). Turkey, in particular, has a mediterranean climate prone to widely varying rainfall. Türkeş and Tatlı (2009) found that the SPI underestimated the extremes, both droughts and pluvials, and suggested a rather involved modification of the SPI method to compensate.

Generally, there is agreement that the most favorable attributes of the SPI are its ability to depict a time-scale (aggregation period) of the user's choice, and that the detection of the start and end of a wet or dry period is straightforward. Along with the wide acceptance of the SPI, it has received relatively little criticism. It is interesting that some workers criticize the use of percent of mean as a measure of wetness or dryness because the mean is not necessarily spatially or temporally consistent as a measure of local climate, yet embrace the SPI believing that it truly normalizes its data because it is rendered with a standardized distribution (e.g., Bordi et al., 2001). Since, in its essence, this standardization is accomplished by dividing the departure from the mean by the standard deviation, there is no more support for the legitimacy of the SPI as a comparable index than there is for percent of mean.

In a study of drought intensity in India, Naresh Kumar et al. (2009) found that the SPI underestimated drought severity when rainfall was extremely low, and underestimated pluvial intensity when rainfall was extremely high. This suggests that the SPI may be less effective in judging the tails of the wet/dry distribution, or that the severity categories (Table 7, Fig. 10) need to be refined. Interestingly, the standard of comparison used by these workers was the deficit index (3). Similar conclusions about the unsuitability of the SPI were drawn by Roudier and

Mahe (2010) in their study of the Bani River in Mali. However, theirs was a somewhat unorthodox application, using daily data with a ten-day accumulation window. This was motivated by the need to depict very-short-duration droughts, typically 10-50 days, characteristic of Mali.

One of the features of the index open to debate is the underlying probability distribution for precipitation. McKee et al. (1993, 1995) recommend the gamma distribution. Guttman (1999) carried out a comparative analysis of five different skewed distributions using data from over 1,000 stations across the conterminous U.S. with more than 65 years of data, from the National Drought Atlas CD-ROM (Willeke et al., 1994). The five distributions were the gamma, Pearson III, generalized extreme value, kappa, and Wakeby. All of these were fitted by a procedure employing L-moments (Guttman, 1993). Guttman also included a second gamma distribution fitted by the method of McKee et al. (1995). Comparison is difficult because, as Guttman notes, there is no standard perfect distribution, so the assessment of the various distributions had to be based on a rather involved process of ranking, for example according to the differences in numbers of dry events per 100 years. Guttman's conclusion is that the Pearson III distribution is the best selection. For present purposes, the more important conclusion is that substantial differences occur in the number of wet and dry periods that the SPI determines due to the selection of the underlying probability distribution and the method of fitting the distribution to data.

Those named above are not the only candidate skewed distributions. Lloyd-Hughes and Saunders (2002) evaluated the log-normal. They also observed that as the time scale increases (i.e., there are more data points in the data aggregation) one might expect the distribution to converge to the normal (consistent with the central limit theorem). Using data from stations across Europe, they tested the suitability of a Gaussian distribution for various windows, finding that for 12 months and greater, the Gaussian distribution was satisfactory. This, of course, greatly simplifies the calculation of the SPI, and for the 1-month timescale, this version of the SPI reduces to the standardized anomaly index discussed above. Zhang et al. (2009) found the log-normal and gamma distributions to perform equally well in their study of the Pearl Basin in China, and used the log-normal. Nalbantis and Tsakiris (2009) in their application to Greece simply employed the log-normal because of its simplicity compared to the two-parameter gamma.

Guttman (1999) also concluded that at least 60 years of data are necessary to achieve an acceptable fit of the central segment of the distribution and 70-80 years for the tails of the distribution. This motivated Wu et al. (2005) to carry out a rigorous study of the sensitivity of the SPI to the length of record. Using data from Nebraska, the SPI based upon fitting the gamma distribution to a long period of record (~90-110 years) was found to be highly correlated ($r > 0.9$) with those fitted from 30-year periods at all time scales from 3 months to 3 years. However, the identification and intensity of drought events displayed more disagreement, becoming greater at longer time scales. The primary source of discrepancy was determined to be differences in the fitted gamma distributions. While a period of record of 30 years is not inappropriate, these authors warn that, “One should be aware that the reduced effective sample size leads to instability of the parameter estimates. If the parameter estimates have little confidence, then the resulting SPI values will also have little confidence.”

There is an additional fly in the ointment. The SPI procedure for fitting the gamma distribution requires $\log S_m$, which is undefined for $S_m = 0$. A modification is necessary if there are zero-valued cumulatives in the data set. The zero values are excluded from the ogive, the gamma function is fitted, and the zeroes are added to the resulting cumulative frequency function. If the proportion of zero values in the set of cumulatives is q, then the $(1\text{-}q)$ nonzero values of S_m are used to construct an ogive to which is fitted the gamma function $G(S)$, see equation (8), and the complete fitted ogive then becomes

$$H(S) \;=\; q + (1\text{-}q)\, G(S) \tag{11}$$

from which the SPI is computed as $\mathcal{N}^{-1}\{H(S)\}$. Typically, the proportion q of zero values of S is small, often zero, so the inverse Gaussian is not problematic. For short time scales in semi-arid climates, or time scales less than 1 year in arid climates, the value of q can become appreciable, and the resulting SPI data points fail to follow a normal distribution. This is not a desirable result for a transform procedure whose goal is to normalize a distribution. The practical effect is that the SPI fails to identify some dry periods, and underestimates the intensity of some of those that it does identify. The problem is exacerbated for short data records. Wu et al. (2007) have carried out an extensive study of this problem for time scales less than 24 weeks, and urge caution in the use of the SPI in these climates. This, of course, includes a substantial part of Texas.

Wu et al. (2001) describe a related index, which has been used in China since the mid-1990's, that they refer to as the China-Z index (CZI). Details of this index are sketchy. It is based upon an underlying Pearson Type III distribution of monthly rainfall standardized by a cube root transformation

$$CZI_i = \frac{6}{C}\left(\frac{C}{2} z_i + 1\right)^{1/3} - \frac{6}{C} + \frac{C}{6} \qquad (12)$$

in which C is the skewness

$$C = \frac{1}{N\sigma^3} \sum_{i=1}^{N} (P_i - \overline{P})^3$$

and z_i is the standardized variate $(P_i - \overline{P})/\sigma$. Wu et al. (2001) generalized this index analogously to the SPI, so that they could evaluate it for different aggregations of the P_i time series, e.g., 3-, 6-, 12-, 24-months etc., thereby effecting a comparison with the SPI for the same time scales. They found the two indices, as well as z_i, to be very similar, except under dry conditions, when the CZI reaches lower (more negative) values than the SPI. They recommend the CZI over the SPI for this reason and because it is simpler to compute.

3.1.3 Palmer index

Despite its common name and its widespread use in monitoring drought conditions, the Palmer index – which is in fact several indices – is a general parameter for tracking moisture availability at the surface of the ground, including both wet and dry conditions. The most familiar index is referred to as the Palmer drought severity index (PDSI). This index has become more widely known in the past couple of decades because of its use as a drought index for large-scale climatological studies, as well as in other disciplines such as dendrochronology and archaeology, and, most recently, because of its central rôle in the products of the Drought Monitor.

3.1.3.1 Formulation

Detailed documentation on the Palmer index historically has been scattered and fragmentary. For many years, Palmer's (1965) report was not widely available, though it is now accessible on the NCDC website. Moreover, the PDSI as computed by NCDC has been modifed several times from the original formulation. Summaries of the method have been presented by Alley (1984), Karl (1983), Guttman (1991), Weber and Nkemdirim (1998), Dai (2011), van der Schrier et al. (2011), and others. There are some specific features of the PDSI of pertinence to the present review not brought out in any of these summaries, however, so a brief review of the methodology is warranted.

The basic indicator is the precipitation anomaly given by (5) above, referencing precipitation P_i over some fixed time increment Δt to $\hat{P}$, a hydroclimatological reference value. The complexity of the PDSI, its physical basis, and its properties derive in part from the definition of the hydroclimatological reference value, which Palmer (1965) refers to as the precipitation *climatically appropriate for existing conditions* (CAFEC). Palmer, whose main concern was agriculture, argues that $\hat{P}$ should incorporate not only the departure from "normal" precipitation for a region, but also precipitation needed to allow the soil to supply its normal vegetational demands, and therefore should include soil-water deficiencies driven by evaporation and transpiration, as well as rainfall deficit.

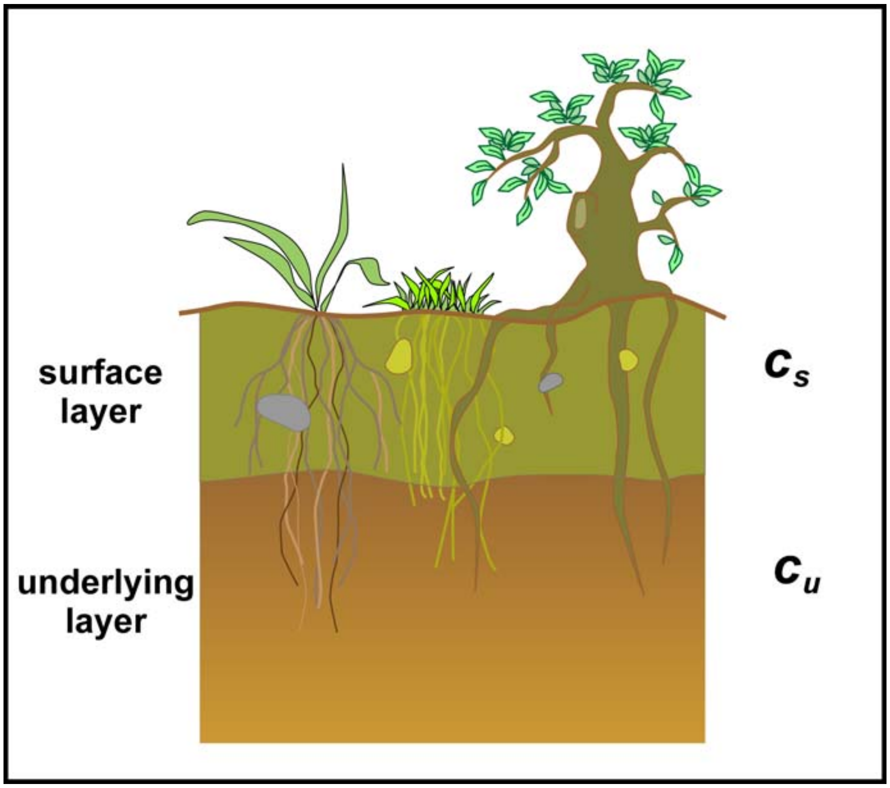

Figure 15 - Soil layers in Palmer soil water-budget model

An elementary water budget is formulated for the upper two layers of soil, Figure 15:

$$\frac{dw}{dt} = p - et - ro - g \tag{13}$$

in which $w = w_s + w_u$ is the total water contained in both soil layers, p, et, and ro are the precipitation, evapotranspiration and runoff rates at the surface, respectively, and g is the water infiltrated to deeper soil layers and aquifers, which is not further considered. As usual, the rates (in depth per unit time) are assumed to be aggregated over some sampling interval Δt, so we define P_i , ET_i and RO_i to be the corresponding cumulative depths of water over the time interval $(i\text{-}1)\Delta t$ to $i\Delta t$, e.g., $P_i \equiv \int_{(i-1)\Delta t}^{i\Delta t} p(t)dt$, and (13) becomes

$$\Delta w_i = w_i - w_{i-1} = P_i - ET_i - RO_i \tag{14}$$

where $w_i \equiv w(i\Delta t)$. Palmer (1965) took Δt to be one month, the conventional accounting interval for long-term climatological studies, but theoretically any time increment could be employed. The soil layers are not characterized, other than by specification of their water-retaining capacities c_s and c_u. Qualitatively, Palmer identifies the upper with the plowed layer and the lower with the root zone. He takes c_s to be 1 inch (2.5 cm), so the depth of the surface layer is implicitly whatever is required to result in a water capacity of 1 inch. The magnitude of c_u is associated in some opaque way with the composition and permeability of the soil.

Although (14) is the conceptual model for the soil water budget, this budget is expressed as accounting rules for the computation of each of Δw_s and Δw_u, as follows:

$$\Delta w_s = \begin{cases} \min\left\{ \begin{matrix} P - PE \\ c_s - w_{s_{i-1}} \end{matrix} \right\} & P - PE > 0 \\ \max\left\{ \begin{matrix} P - PE \\ - w_{s_{i-1}} \end{matrix} \right\} & P - PE < 0 \end{cases} \tag{15a}$$

$$\Delta w_u = \begin{cases} \min\left\{ \begin{matrix} P - PE - \Delta w_s \\ c_u - w_{u_{i-1}} \end{matrix} \right\} & w_{s_i} = c_s \\ 0 & c_s > w_{s_i} > 0 \\ \max\left\{ \begin{matrix} (P - PE - \Delta w_s)\ \Theta \\ - w_{u_{i-1}} \end{matrix} \right\} & w_{s_i} = 0 \end{cases} \tag{15b}$$

where $\Theta \equiv w_{u_{i-1}} / (c_s + c_u)$. In these equations, all quantities are assumed to be evaluated at $t = i\Delta t$, so the subscript i is omitted unless the previous time step $i-1$ is intended, as in the definition of Θ. These accounting rules embody Palmer's assumptions that evapotranspiration from the surface layer proceeds at the rate of potential evapotranspiration PE (discussed below), that water additions to the lower layer occur only after the upper layer is at capacity, and that water losses from the lower layer occur only after the upper layer has zero water. Although the lower layer is allowed to reach capacity governed only by the surfeit of water at the surface, the factor Θ prevents the soil water from zeroing however large the surface deficit, but rather forces it to converge asymptotically to zero.

Once the values of Δw_s and Δw_u are available, the additional terms in (14) may be readily calculated:

$$ET = \begin{cases} PE & P - PE \geq 0 \\ P + \max\{0, -\Delta w\} & P - PE < 0 \end{cases} \tag{16}$$

$$RO = \max\{0, P - PE - \Delta w\} \tag{17}$$

in which $\Delta w = \Delta w_s + \Delta w_u$. Palmer (1965) prefers to distinguish between positive and negative values of Δw, referring to the former as "recharge" and the latter as "loss" (perhaps due to the asymmetry of Δw_u arising from the Θ factor). Thus, he defines recharge $R \equiv \max\{0, \Delta w\}$ and loss $L \equiv \max\{0, -\Delta w\}$.

Potential evapotranspiration, the rate of water transfer to the atmosphere from a homogeneous parcel of vegetated land with saturated soils, is a key parameter in the Palmer soil-water budget. The term was introduced by C. W. Thornthwaite in 1944 (Willmott et al., 1985), and even though it is an imperfect concept (Brutsaert, 2005), it has received considerable currency among climatologists. Thornthwaite (1948) devised an empirical equation for its computation requiring only monthly temperature data:

$$PE_j = C\,(10\,T_j / I)^a\,(n/30)\,(d/12) \tag{18}$$

where T_j = temperature (°C) for month j (j = 1 … 12)

I = annual heat index given by:

$$I = \sum_{j=1}^{12} I_j \quad \text{where } I_j = (\max(0, T_j\}/5)^{1.514}$$

a = $0.49239 + 0.017921\, I - 7.71\text{x}10^{-5}\, I^2 + 6.75\text{x}10^{-7}\, I^3$

n = number of days in month j

d = day length (hrs) for month j

= $2\, H/15$ where H = the hour angle in degrees:

$\cos H = -\tan\varphi \tan\{23.45 \cos[\,(2\pi/365)\ 173/N_j]\}$

φ = latitude

N_j = day number (a.k.a. Julian day) of the middle of month j

C = 1.6 cm/mo

Palmer (1965) adopted the Thornthwaite equation for potential evapotranspiration. This is a monthly equation, so to proceed further in summarizing Palmer's (1965) development, we take the time interval Δt to be a month.

For the computational machinery available in the mid-twentieth century, equation (18) posed substantial difficulties, so it was implemented by various nomograms and tables (e.g., Thornthwaite and Mather, 1955, 1957; van Hylckama, 1959), one of which was contributed by Palmer and Havens (1958). The original intent was for application in climatology, so that the monthly indexes I_j and the annual index I would be evaluated from historical records or climate normals and would be fixed for a given location. However, the method was embraced as an expedient means for computing PET, or sometimes estimating actual evapotranspiration, for which the monthly and annual heat indexes could be considered time variables as well, evaluated either for a given calendar year, or over the preceding year ending in the month of interest (van Hylckama, 1959; Sellinger, 1996; Xu and Singh, 2001; Fisher et al., 2011; van der Schrier et al., 2011).

As matters developed, the Palmer index proved to be relatively insensitive to the details of PET so these distinctions are of minor import. Karl (1986a) reports an experimental calculation in which he used a fixed annual cycle of monthly PET in a 53-year computation of the PDSI, and found negligible effects on the PDSI. Mavromatis (2007) substituted the Priestly-Taylor PET for Thornthwaite, but found no improvement in the performance of the PDSI. Van der Schrier et al. (2011) compare the effects of the Thornthwaite and Penman-Monteith formulae for PET. Even though these alternative methods were found to have "vast" differences, the net effect on the PDSI was minor, due to the various normalizations and scalings in the PDSI, and because precipitation generally dominates the calculation.

In addition to potential evapotranspiration, Palmer (1965) introduced three more "potential" quantities, *viz*. potential recharge

$$PR = c - w_{i\text{-}1} \tag{19}$$

potential loss

$$PL = \min\{w_{s_{i\text{-}1}}, PE\} + (PE - \min\{w_{s_{i\text{-}1}}, PE\})\ \Theta \tag{20}$$

and potential runoff

$$PRO = c - PR \tag{21}$$

While (19) and (20) are plausible, (21) is decidedly peculiar. Palmer was led to this formulation by postulating that *PRO* must depend in someway on soil moisture. If "potential precipitation" *PP* is the amount necessary to supply potential recharge and potential runoff in the absence of evapotranspiration, i.e., $PP = PR + PRO$, then (21) implies $PP = c$, a result not "particularly elegant" remarked Palmer (1965), but, "It has worked out better than expected." The fallacy is that *PRO* must also depend on precipitation, which is not incorporated in (21). As Palmer anticipated, a simple large constant value assigned to *PP* (or, put another way, a large constant added to c) will entail a more satisfactory behavior, for example in removing the occasional occurrences of negative values of *PP*. (Other than exposing an apparent inconsistency in the formulation, *PP* plays no rôle in the PDSI.) The form (21) is retained in the present analysis to be consistent with the original PDSI.

Palmer (1965) postulated that, for each month, the ratio of the long-term average value of each of the water-budget variables to the long-term average of its corresponding potential value is a fixed constant, i.e.,

$$\alpha_j \equiv \frac{\overline{ET}}{\overline{PE}} \qquad \beta_j \equiv \frac{\overline{R}}{\overline{PR}} \qquad \gamma_j \equiv \frac{\overline{RO}}{\overline{PRO}} \qquad \delta_j \equiv \frac{\overline{L}}{\overline{PL}} \tag{22}$$

are characteristic monthly values, $j = 1, \ldots, 12$, for a given location. Here the overbar is a long-term average of the values for month j, $\overline{X} = (\sum_{i \equiv j (\mathrm{mod}\, 12)} X_i) / n_j$, n_j denoting the number of data for month j in the meteorological time series. More importantly, it is these coefficients that define Palmer's CAFEC values. For each month i in the time series, the CAFEC parameter is defined to be the product of its respective coefficient and its potential value for that month, e.g., $\langle ET \rangle_i \equiv \alpha_j PE_i$, $\langle RO \rangle_i \equiv \gamma_j PRO_i$, etc., where j is the calendar month of i, that is, $i \equiv j$ (mod 12), $j = 1$ denoting January *et seq*. We note that this definition implies that $\overline{X} = \overline{\langle X \rangle}$, for each of the water-budget variables X. Finally, in analogy to the soil-water budget, CAFEC precipitation is defined to be:

$$\hat{P}_i \equiv \langle ET \rangle_i + \langle R \rangle_i + \langle RO \rangle_i - \langle L \rangle_i \tag{23}$$

The complexity of (5) in the PDSI is now apparent: not only is its computation based on a time series of a complicated soil-water budget, but the reference value $\hat{P}$ for the moisture anomaly is itself a time series, with monthly values dependent upon the "potential" values of the terms in the soil water budget.*

Palmer (1965) exemplified the utility of $\hat{P}$ as an indicator by using the Dust Bowl years in western Kansas, specifically June 1932 - May 1935, for which the cumulative rainfall was 40.7

* Palmer (1965) tabulates monthly values for the Dust Bowl years of all of the above water-budget terms for central Iowa, and of the potential and CAFEC terms for western Kansas. However, some frustration will result from attempting to match Palmer's numbers exactly because the NCDC divisional monthly means have been modified since Palmer's work, both to correct errors and to compensate for varying observation times (see Karl et al, 1986).

Table 8
Palmer (1965) climatological weighting factors K_j for Texas NCDC divisions based on 1931-60 normals

		High Plains	*Edwards Plateau*	*South Texas*
	Division	*1*	*6*	*9*
jan	1	1.95	1.14	1.28
feb	2	2.09	1.31	1.39
mar	3	1.83	1.26	1.60
apr	4	1.66	1.05	1.17
may	5	1.24	0.96	1.17
jun	6	1.32	0.94	1.17
jul	7	1.43	1.13	1.43
aug	8	1.79	1.40	1.28
sep	9	1.48	0.90	1.04
oct	10	1.42	1.00	1.18
nov	11	1.99	1.39	1.57
dec	12	1.79	1.17	1.33

ins. This would be a cumulative deficit of 13.6 ins compared to the cumulative long-term average for this calendar period of 54.3 ins. Relative to the cumulative CAFEC precipitation of 69.1 ins, however, the deficit is 28.4 ins, which reflects not only the below-average rainfall but the effect of unusually high temperatures, hence high PET.

The precipitation anomaly (5) is now multiplied by an empirical scaling (or "weighting") factor to arrive at the Z-index:

$$Z_i = K_j\,(P_i - \hat{P}_i) \quad \text{for } i \equiv j \pmod{12} \tag{24}$$

where

$$K_j = \left(17.67 \Big/ \sum_{m=1}^{12} \overline{D}_m k_m \right) k_j \tag{24a}$$

$$\overline{D}_j = \sum_{i \equiv j (\mathrm{mod} 12)} \left| P_i - \hat{P} \right| / n_j \tag{24b}$$

$$k_j = 1.5 \log_{10} \left\{ \left(\frac{\overline{PE}_j + \overline{R}_j + \overline{RO}_j}{\overline{P}_j + \overline{L}_j} + 2.8 \right) \frac{1}{\overline{D}_j} \right\} + 0.5 \tag{24c}$$

and j = 1, …, 12, with 1 denoting January *et seq*. The purpose of the monthly scaling factor K_j is to "normalize" the moisture anomaly for the climatological region, i.e., to transform the anomalies to a constant scale independent of the local hydroclimatology. By comparing central Iowa and western Kansas, Palmer (1965) first devised a statistical scaling based on the ratio ($\overline{PE}_j + \overline{R}_j$) to ($\overline{P}_j + \overline{L}_j$), then augmented the relation to include a total of nine NCDC divisions in Iowa, Kansas, Tennessee, Ohio, Pennsylvania, North Dakota, and Texas (3). The monthly K factors for Texas are shown in Table 8.

Originally, Palmer experimented with the cumulative scaled moisture anomaly $\sum Z_i$ as an index, and using the 13 driest periods (of varying lengths) in central Iowa and western Kansas as standards, applied an additional scaling to bring these to an (arbitrary) value of –4.0, representing "extreme drought." However, he felt that the cumulative anomaly underrepresented the most recent monthly data in favor of earlier months in the time series. In particular, Palmer (1965) was unsatisfied with a cumulative indicator that continued to be influenced by data from the unlimited past. Instead, he proposed the index

$$X_1 = Z_1/3 \tag{25a}$$

$$X_i = Z_i/3 + c\, X_{i-1} \tag{25b}$$

This recursive prescription is suitable for time-advancing computations, but its mathematical properties are more transparent in its equivalent form

$$X_i = \frac{1}{3}\sum_{n=1}^{i} c^{i-n} Z_n \tag{26}$$

The constant c = 0.897, yet another empirical coefficient, was inferred from scaling of extreme drought to a value of $X_i \approx$ –4.0. The Palmer moisture categories are presented in Table 9.

Overlaid on the index formulation is a strategy of computation in which (25) is *re-initialized* at the beginning of a "dry spell" or "wet spell." This of course derives from Palmer's principal

Table 9
Moisture categories for PDSI,
see Palmer (1965), Kangas and Brown (2007)

designation	PDSI *value**
extremely wet	≥ 4.0
very wet	[3.0, 4.0)
moderately wet	[2.0, 3.0)
slightly wet	[1.0, 2.0)
incipient wet spell	[0.5, 1.0)
near normal	(-0.5, 0.5)
incipient drought	(-1.0, -0.5]
mild drought	(-2.0, -1.0]
moderate drought	(-3.0, -2.0]
severe drought	(-4.0, -3.0]
extreme drought	≤ -4.0

* $(a,b]$ designates a numerical interval inclusive on the right, that is, all values x such that $a < x \leq b$. $[a,b)$ denotes an interval inclusive on the left.

interest in agricultural impacts of drought, in which the water stored in soil is of greater importance than the antecedent cumulative surplus of rainfall (most of which is presumed to have run off), and led Palmer into the murky issue of identifying the beginning and end of a period of abnormal moisture (either wet or dry). He was motivated by two concerns. First, agricultural impacts of a short dry period (two or three months) after a month of heavy rainfall appeared underestimated by the raw value of the index (25). Second, any rule to initiate or terminate a wet or dry period must consider how (25) varies over the succeeding, i.e., future, data in the time series. Palmer ultimately evolved a complex, threshold-based procedure that requires tracking three versions of (25) separately, corresponding to potential ongoing wet period, potential ongoing dry period, and an established ongoing wet or dry period. Choice among these is based upon values of the water-budget terms over the entire duration of the wet or dry period, which for most of the months in that period includes both past and future values. Palmer (1965) assigned a trigger value of X_i = -1.0 or X_i = +1.0 for the start of a dry or wet period, resp., whereupon equation (25a) was applied to re-initiate the computation of X_i. Palmer assigned a trigger value of X_i = -0.5 or X_i = +0.5 for termination of a dry or wet period, resp., subject to additional rules involving their occurrence in the time period and the accumulated moisture implicit in the time series for Z_i in (24). Details of this procedure are given in Palmer (1965), and more explicitly in Karl (1983, 1986b), Alley (1984), Karl et al. (1987) and Guttman (1991). The index (25) whose values depend upon the rules for re-initiating the index, i.e., the pattern of wet spells and dry spells, is the PDSI.

3.1.3.2 Properties and application

The Palmer index, particularly the PDSI, was a monumental step forward in the analysis and monitoring of drought in the United States and elsewhere. The NCDC has provided the PDSI by climatic division as a routine product for many years. The original baseline period, or "calibration period," was the 1931-60 standard climatic normal. According to Karl (1986b), the present calibration period is 1931-90. Clearly, the PDSI is a parameter designed for climatological evaluation, since it relies upon a complete time series being available with both "past" and "future" values of the water-budget terms for each month in the time series. A closely related index has been formulated by NCDC in which the decision process for terminating a drought is simplified so as to eliminate need for the "future" values of the water-budget time series, but is instead determined entirely from "past" values (Heddinghaus and Sabol, 1991; Karl, 1986b; Guttman, 1991). This index, the Palmer Hydrological Drought Index (PHDI), is identical to the PDSI, excepting identification of dry and wet periods. More importantly it is operational, in that it can be computed in real time because it does not require "future" values for its calculation. NCDC provides updated monthly values of the Z index (12), the PHDI and the PDSI in its data product file TD-9640, see Section 2.1.2 above.*

The Palmer index is neither a pure meteorological index nor a hydrological index, but intermediate between the two. It is more than a meteorological index (like those reviewed in Sections 3.1.1 and 3.1.2) because its formulation includes a water budget of the upper layers of soils, albeit rudimentary. But, as Alley (1985) showed, it is not a hydrological index either, because such an index as a statistic of streamflow, lowest quartile for example, behaves

* TD 9640 contains a fourth Palmer index, the modified PDSI, which does not differ substantively from the PDSI in its utility as a hydroclimatological indicator and is not further considered in this report.

Table 10
Restarts of PDSI for Texas divisions, 1895-2011

	Division	*restarts*	*frequency (per year)*		*Division*	*restarts*	*frequency (per year)*
1	Hi Plains	157	1.34	6	Edwards	144	1.23
2	Lo Plains	155	1.32	7	S Cent	129	1.10
3	N Cent	141	1.21	8	Upr Coast	159	1.36
4	East	157	1.34	9	South	150	1.28
5	Trans-Pecos	154	1.32	10	Valley	155	1.32

differently than the PDSI, one frequently indicating drought when the other does not. This conclusion could perhaps have been anticipated because the Palmer indices focus primarily on those terms (processes) of the soil-water budget, of which runoff is ancillary, while runoff is the principal process driving streamflow.

The procedure of re-initializing the index (25) at the beginning of a dry or wet spell has two effects. The first is to completely alter the utility of the drought categories (Table 9) as thresholds of moisture conditions. The second, and perhaps more important, is to disrupt the progression of the response of the index to the changing soil water budget. This was indeed the objective of Palmer, finding that the index (25) did not seem to respond quickly enough to changing moisture conditions to satisfy himself. Of the indices reviewed in this study, this procedure is unique. The usual strategy is to define the index however appropriate, then use thresholds to categorize the resultant moisture conditions, including anomalous periods. No other index is modified in the course of determining the beginning or end of a wet or dry spell. There are numerous such restarts in the PDSI time series. In the Texas divisions, these average more than once a year over the 1895-2011 period, see Table 10.

Something can be learned about the time response of the Palmer index from its mathematical formulation. The closed-form version of the moisture index (26) is written as a convolution, and comparison with (1) implies that the kernel of the Palmer index operating on the Z-index time series (24) is:

$$h_i = \frac{1}{3}\, c^{\,i-1} \qquad\qquad i = 1, 2, \ldots, N$$

in which $c = 0.897$ and N is the number of data points in the time series. This is an exponential-decay impulse response with a time constant of 9.20 months. The effect of Z values older than about two time constants (18.4 months) would be reduced by about 90%, and for three time constants (27.6 months) would be virtually eliminated. This is in contrast to the spectral analyses of Guttman (1998) in which he interpreted the memory of the PDI, with wet/dry period protocols, to be 2.5 to 7 years, depending upon the spectral type.

Guttman (1991) carried out a study of time response in which the Palmer model for PHDI, again with wet/dry period protocols, was driven by cycling inputs of annual long-term mean temperature and precipitation until the model equilibrated at PHDI = 0. He discovered, first, that the PHDI is subject to a "starting transient" caused by inconsistencies or errors in the initial conditions, arising from the arbitrary nature of the initial values of soil water and moisture anomaly (24). Some (simulated) time is needed for these starting transients to flush out of the solution. This time, it turns out, depends upon the aridity of the site, and ranges from a few months at humid sites to as much as four years at arid sites. Guttman (1991) recommends that the first four years of any long-term PDI simulation be ignored. After the simulated index equilibrated, that is, about half way through his simulation period, Guttman introduced a one-month perturbation in either precipitation or temperature and determined the response of the model. Temperature perturbation was found to be "practically insignificant" in engendering a response in the PDI. Precipitation was a different matter. Even a 25-50% change in precipitation induced an incipient or established wet or dry spell, corresponding to the sign of the perturbation, whose duration ranged from 6 to 24 months. An eyeball-estimate from Guttman's Figure 9 indicates a time-constant of about 18 months in the response, about twice the time constant determined from the mathematical method above. The mathematical result is strictly applicable only to (25) without the re-initialization protocols.

There have been several studies of droughts and pluvials in the U.S. based upon the Palmer index. Diaz (1983) analyzed historical dry and wet episodes in the period 1895-1981, in which a period of at least three months with PI <-2.0 is defined as a dry episode and with PI > 2.0 a wet episode. A similar period of at least six months is a "major" episode. The PDSI played a central rôle in the formulation of a climate extremes index used by Karl et al. (1996) to examine whether the U.S. climate has become more extreme since the 1970's. Fye et al. (2003) used the PSDI as the basis for examining twentieth-century pluvials and droughts. Kangas and Brown (2007) performed a study of pluvials and droughts over the contiguous U.S. using the PDSI based upon the 4-km gridded interpolated data set from PRISM, see Section 2.1.2, and available water capacity data from an extensive soil properties data base at Pennsylvania State University. Noting that the Palmer index has a timescale of 6-12 months (which agrees with the above time-response analysis), these authors compare the PDSI with the SPI-6 and SPI-12, finding that they match well overall, particularly in identifying droughts and pluvials, though they randomly differ in the intensity of the events. PDSI was found to be intermediate between SPI-6 and SPI-12, inconsistently agreeing with one or the other. None of the drought indices used by Kangas and Brown identified 1905-1917 as a major pluvial, as reported by Fye et al. (2003), but did identify the increase in wet events since 1970 found by Karl et al. (1996). Among many regional studies using the PDSI or one of its variants, Dahm et al. (2003) used several Palmer indices to study the effect of drought on tributaries of the Rio Grande in central New Mexico, including the impacts on geochemistry of these streams. In the period 1895 to 2000, two extended droughts were identified, 1899–1904 and 1950–56, as well as several intense but much shorter droughts. These authors present associated impacts on riparian and aquatic species and their association with the degree of drought indicated by the categories of the Palmer indices (Table 9).

While the acceptance of the PDSI in other countries has not been as extensive as in the U.S., it has been applied to Europe (Briffa et al., 1994; Lloyd-Hughes and Saunders, 2002). Dai et al. (2004) and Dai (2011) used a global data set of PDSI values using the Penman-Monteith equation for potential evapotranspiration to study the worldwide distribution of drought. Dai (2011) found the basin-averaged values of the PDSI (and the self-calibrated PDSI, see below) to be correlated with annual flows of most of the top 230 rivers of the world (correlations ranging 0.4 to 0.9). He also found good correlation of the PDSI with soil moisture. Vicente-Serrano et al. (2011) report a similar evaluation and found similar correlation of annual precipitation and SPEI (Section 3.1.4.3) with annual flow, opining that it is the strong dependency of PDSI on precipitation that entails the hydrological correlation.

The performance of several of the Palmer indices is presented in Figures 16-23 for selected periods and climatic divisions in Texas. These are graphs of the monthly division-mean indices from the NCDC files TD-9640 of the Z-index, equation (24), the PDSI with the Palmer protocols for identifying wet- and dry-periods, i.e. the re-initialization method, and the PHDI with the NCDC protocols. In addition, the Palmer index X, equation (25) *without re-starts*, is plotted as well. The Z-index, it will be recalled, responds primarily to precipitation and secondarily to temperature, through the effect of temperature on water demand as measured by potential evapotranspiration. The only memory that the Z-index has is what is implicit in the soil water budget (and the autocorrelation of rainfall and air temperature). The other three indices are integrators of the meteorology.

Figures 16-18 display the five years spanning the end of the Drought of the Fifties. Until early 1957, the three divisions shown were suffering from this prolonged drought. The PDSI, PHDI, and X indices all track together (see "A" marked on these figures). This is characteristic of a prolonged drought, and is not disrupted until there is sufficient rainfall to trigger the dry-spell-termination protocols. In response to the May rainfall in 1957, the PDSI shifts suddenly from a -5 drought to a pluvial that in two months climbs over +4 in both the North Central (Fig. 16) and Edwards Plateau divisions (Fig. 17) and nearly this much in South Texas (Fig. 18), see "B". Both the PHDI and X index have a slower response and do not reach such high positive values. In South Texas (Fig. 18), after three months in the positive range, the PDSI suddenly declares a two-month drought, see "C". This evidently exemplifies the responsiveness that Palmer was trying to achieve in this index. The X index is more moderate in response and does not emerge from the Drought of the Fifties until early 1958 ("D" in Fig. 18). The North Central slides back into drought again by the middle of 1959 ("D" in Fig. 16), though the PDSI falls abruptly into the dry range in summer of 1958 ("C" in Fig. 16). The East Texas Division, Figure 19, was not as affected by the drought as the rest of the state. In the early years of 1950-54, shown in this figure, drought was more moderate, initially occurring in late 1950 ("A" in Fig. 19). While the PDSI registers a one-year wet period starting in late 1952 ("B"), the X index really never rises out of a dry period. Note the abrupt rises in the PDSI at "B" and "D" in Fig. 19.

Figure 16 - Palmer indices, monthly, North Central Division (3), 1955-59, see text.

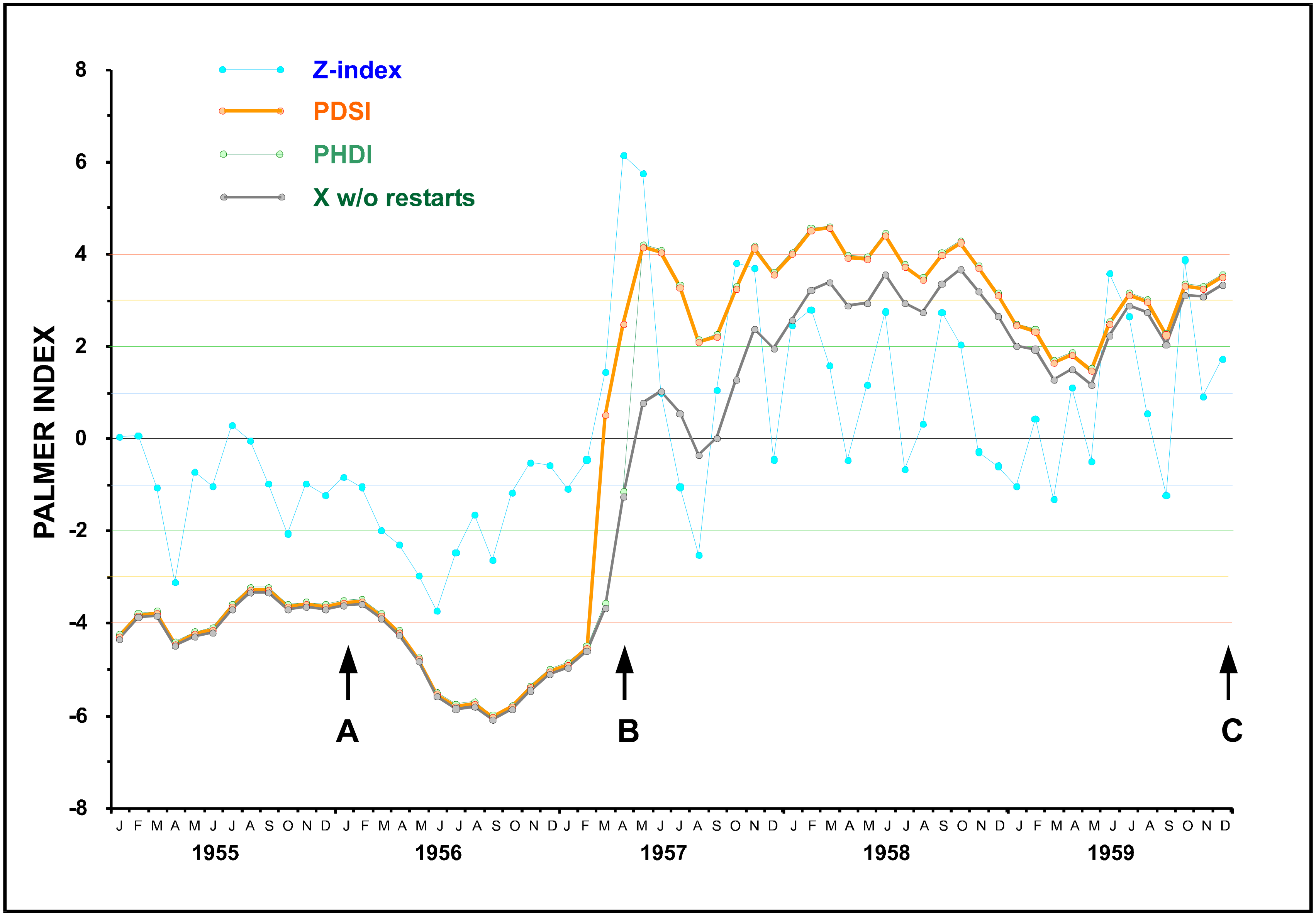

Figure 17 - Palmer indices, monthly, Edwards Plateau Division (6), 1955-59, see text.

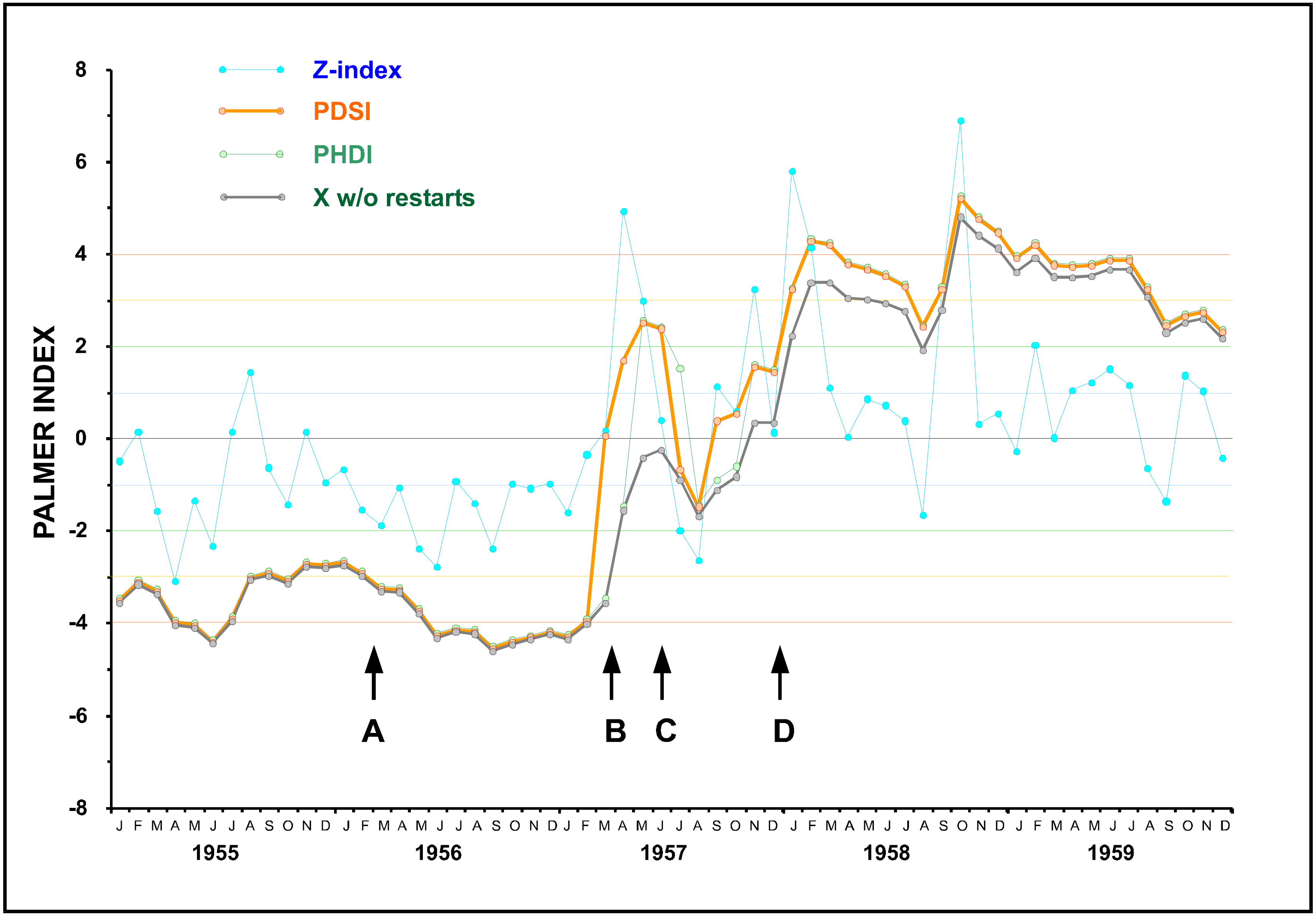

Figure 18 - Palmer indices, monthly, South Texas Division (9), 1955-59, see text.

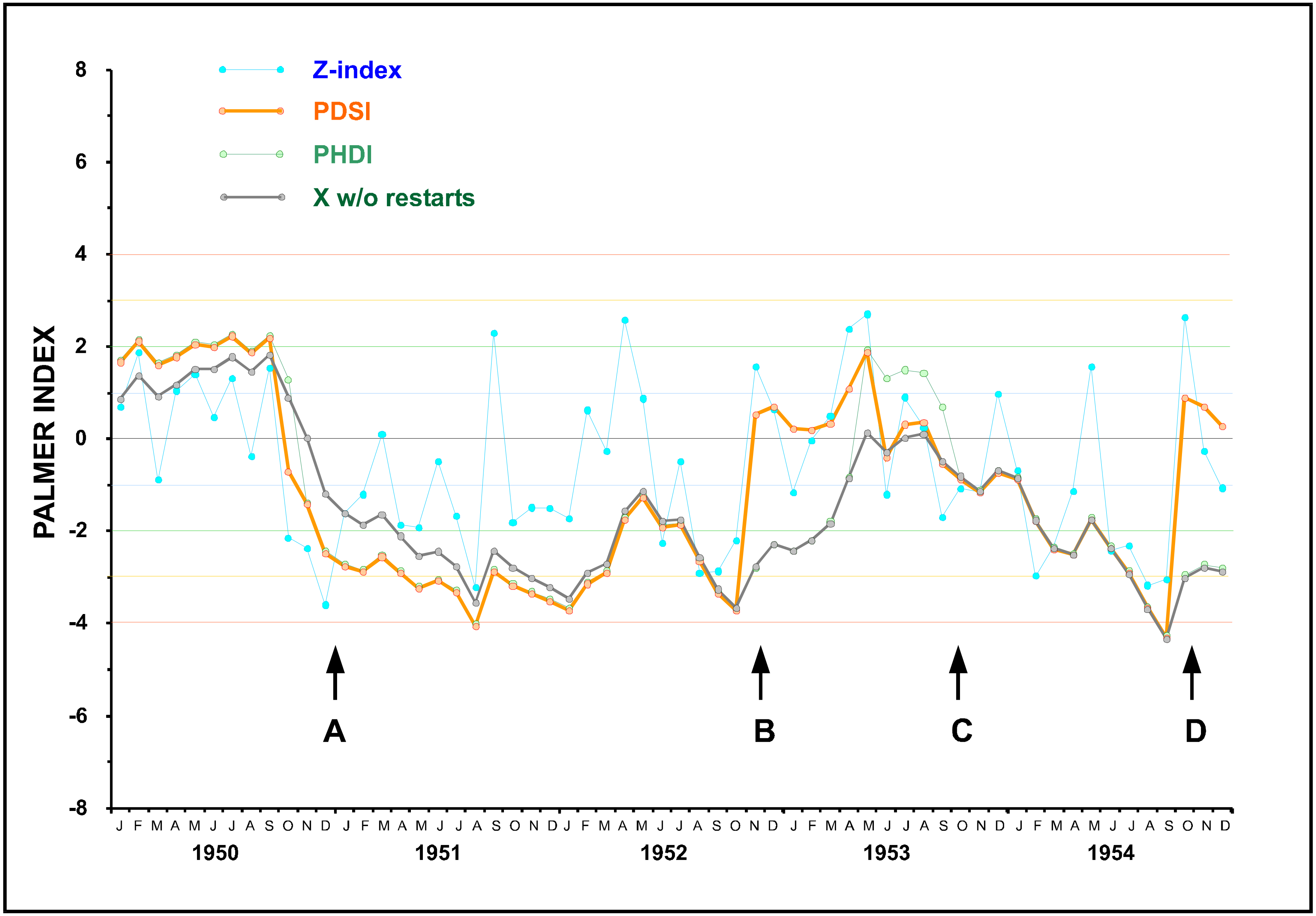

Figure 19 - Palmer indices, monthly, East Texas Division (4), 1950-54, see text.

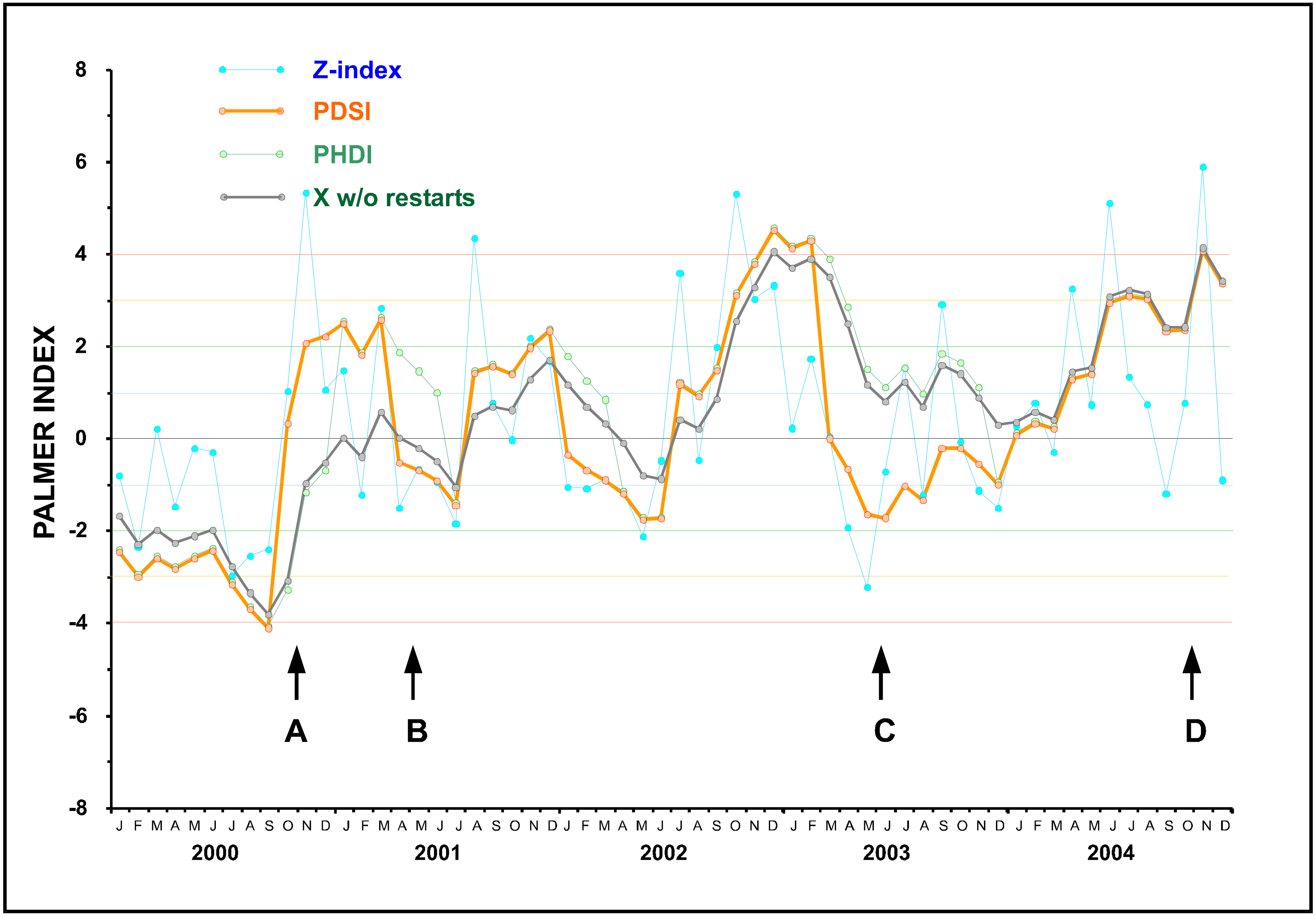

Figure 20 - Palmer indices, monthly, South Central Division (7), 2000-04, see text.

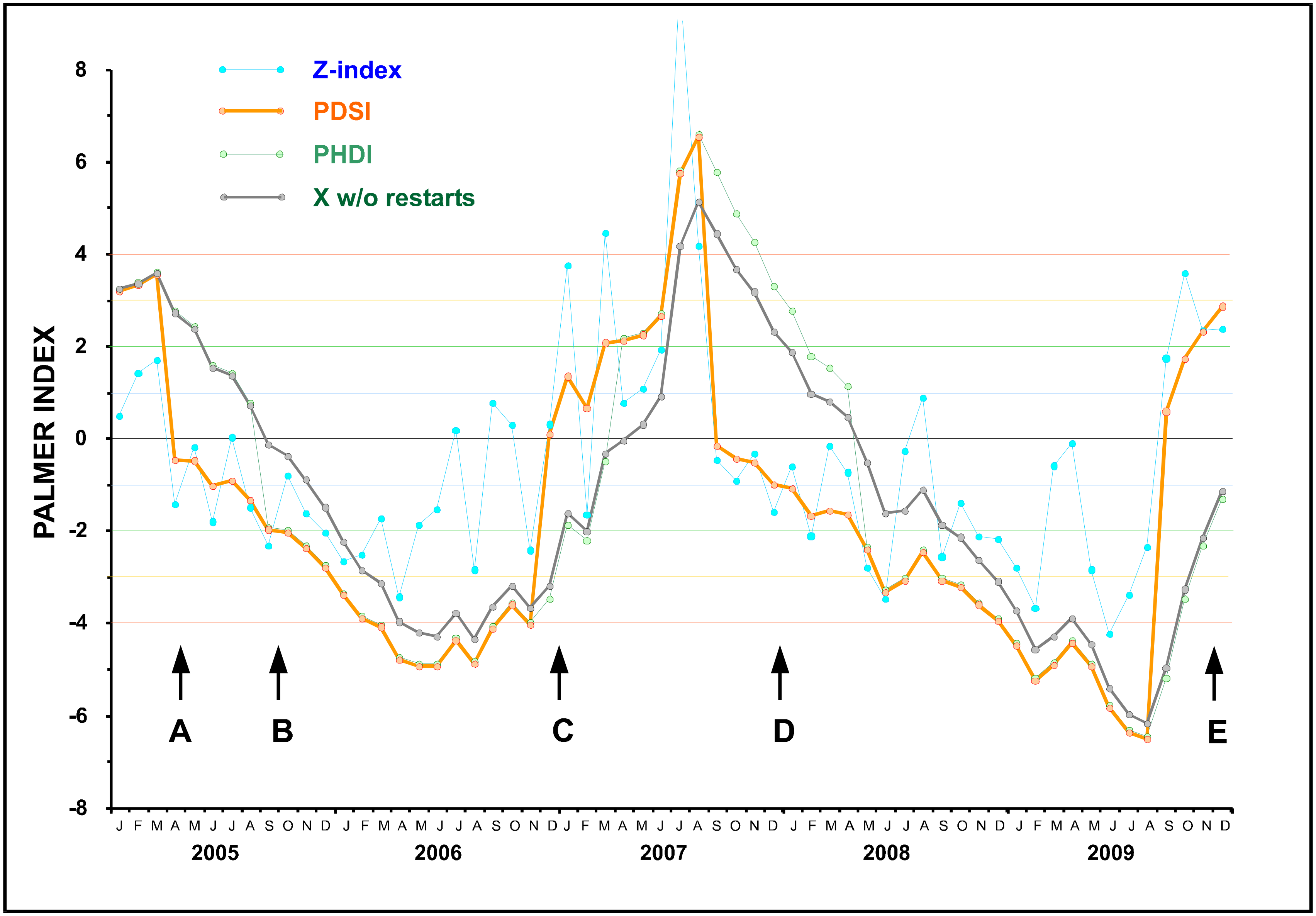

Figure 21 - Palmer indices, monthly, South Central Division (7), 2005-09, see text.

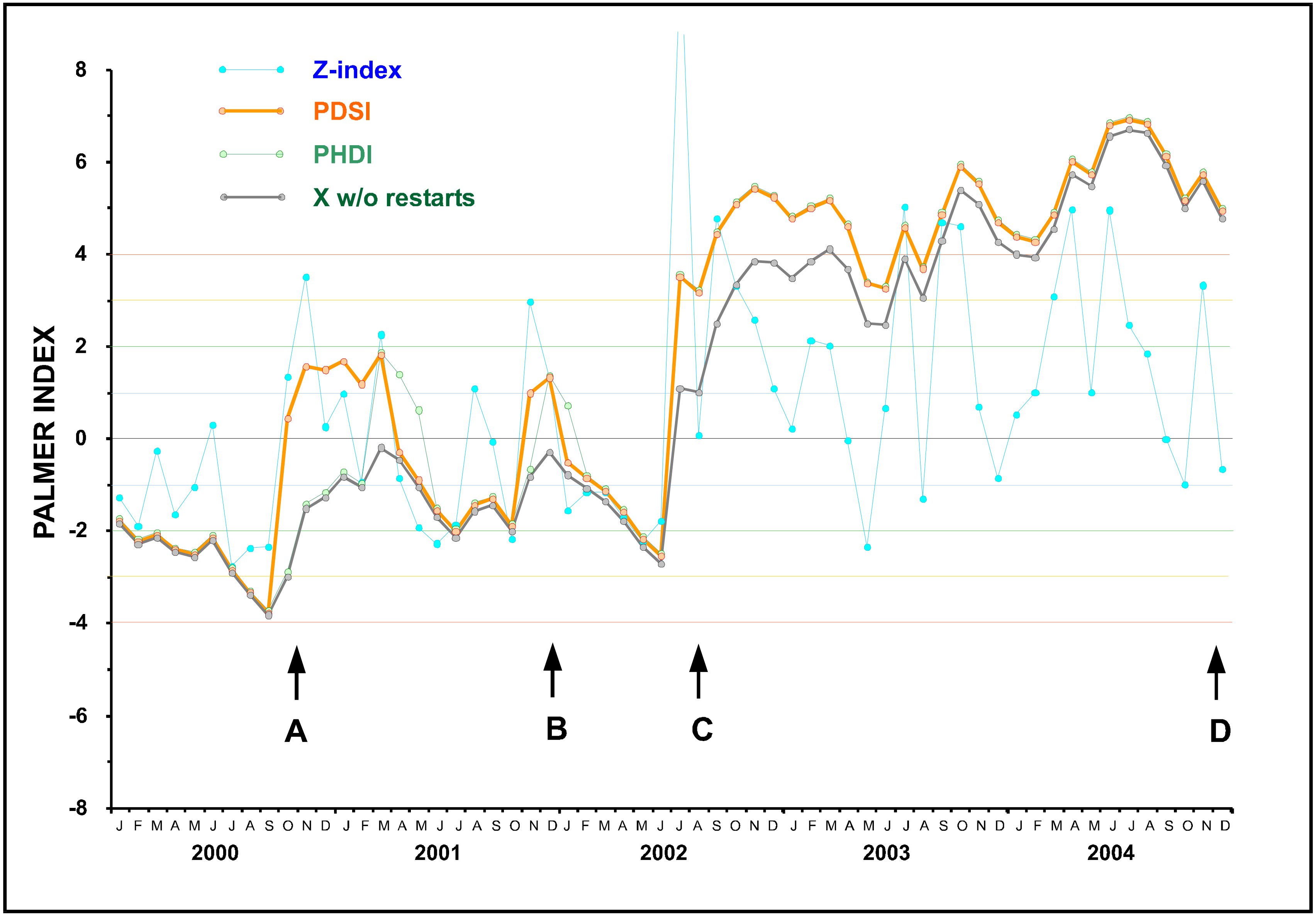

Figure 22 - Palmer indices, monthly, South Texas Division (9), 2000-04, see text.

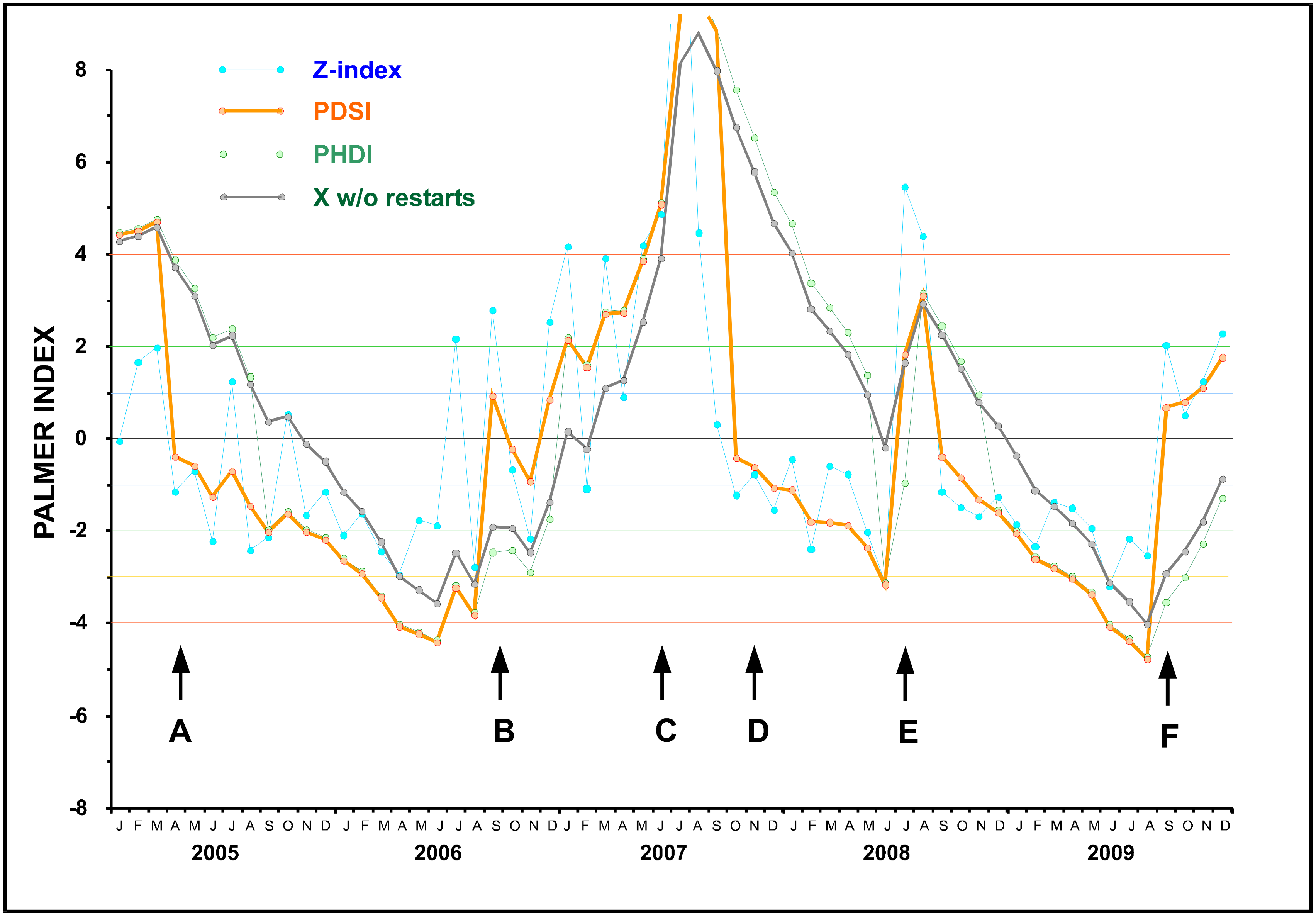

Figure 23 - Palmer indices, monthly, South Texas Division (9), 2005-09, see text.

The ten-year period 2000-2009 was one of the most variable decades on record, and is displayed in Figures 20-23 for the South Central and South Texas divisions. The 2000-04 period was one of increasing moisture conditions from drought to pluvial. In the South Central division, Fig. 20, there is a general upward trend in all three indices, but, as might be expected, the PDSI exhibits much more variability, vacillating from dry periods to wet periods four times during this five-year record. The excursions at "A" and "B" in Fig. 20 should be especially noted, in comparison to the more moderate X index. For one year, "C" in the figure, the PDSI registers dry conditions while the PHDI and X indices register wet. In 2005-09, there are two intense droughts separated by a major pluvial. There are several abrupt changes in the PDSI from one moisture condition to another in Fig. 21, in April 2005 (marked "A"), December 2006, September 2007 and August 2009. There are also four periods of substantial duration in which the PDSI and X index indicate opposite moisture conditions ("B", "C", "D", "E"). In South Texas, the change from drought to wet conditions, as indicated by the X index, occurs more suddenly than in the South Central ("C" in Fig. 22), but there is much more vacillation in the PDSI over this period ("A", "B", "C"). For the prolonged pluvial starting in 2002, all three indices track together. In the South Division, the pluvial ("C" in Fig. 23) separating the two intense droughts is the largest on record for this division. In addition to the sudden excursions of the PDSI over this period, this index leads the X index, sometimes by as much as a year. It is clear from this example that the wet/dry-period protocols greatly alter the response of the index. The shift in PDSI at "B", "D", and "F" should be noted. At "D", the index drops over nine points in one month.

Several reviews of the Palmer drought index have appeared in the literature, and some of its deficiencies have been well-identified. Concerned that the persistent droughts in the U.S. interior indicated by the PDSI might be an artifact, Karl (1983) undertook an extensive sensitivity study of the index (see also Karl and Koscielny, 1982). He determined that the index is relatively insensitive to the available water capacity but sensitive to the value of K. However, since K influences moisture-excess and moisture-deficit conditions equally, there is no impact on durations of wet and dry spells, and the interior drought persistence is real. Additional sensitivity studies were reported by Karl (1986b), addressing the response of the index to a change in baseline period. He found substantial changes in the PDSI and PHDI between the 1931-60 normal and the 1951-80, and greater consistency for calibration periods longer than 30 years. The drought index was found to be much less sensitive to whether PE is computed monthly as an annual cycle versus a single annual calculation, see equation (18).

Alley (1984) carried out a thorough review of the Palmer index, finding several deficiencies, one of the most serious to be that the PDSI was developed for a limited part of the Midwest, with a modest extension to encompass nine climatic divisions in seven states. There is little evidence that the K factor, designed to normalize the index for application to other climates, is adequate to the task. Indeed, Palmer's original intent to scale the index to the range -4 to +4 clearly fails for other geographic regions of the U.S., including Texas, where its range is nominally ± 6 and greater (cf. Figs. 16-23). Globally, it ranges ± 10 (Dai, 2011). Guttman et al. (1992) examined the statistical frequency of PDSI drought categories nationwide. They found that "severe" drought ($-3 \geq \text{PDSI} > -4$) occurred at least 5% of the time in most of the country and at least 10% of the time west of the Mississippi, and that "extreme" drought ($-4 \geq \text{PDSI} > -5$) occurred at least 5% of the time in the western states. Although Palmer's (1965) use of the 13 driest events in Kansas and Iowa to calibrate his index did not specifically address frequency of occurrence, Guttman et al. (1992) argue that a "severe" event should occur every 15-20 years and an "extreme" event every 25-50 years, and therefore Palmer's categories exhibit too high a frequency. Moreover, the frequency of these categories should be homogeneous but in fact are highly variable across the country. Guttman's (1998) spectral analysis of the PDSI determined that the shapes of the spectra (out to a period of 108 months) were regionally inconsistent, from which, as noted above, he concludes that the index is difficult to interpret.

Wells et al. (2004) devised a modification of the Palmer index in which a more rigorous calculation of K_j replaces the empirical expression of (24), relating this strictly to specific station data, and in which the coefficient c in (25) is more rigorously defined to scale the drought index to -4.0 and +4.0 at the 2% and 98% exceedance values of X_i in (25), resp. They dubbed this the self-calibrating Palmer index (scPDSI) because it generalizes the determination of PDSI from station parameters and removes the dependency of calibration being based on a specific geographical locale. (A FORTRAN computer for computing the scPDSI is available for download at http://nadss.unl.edu.) This index has been widely applied, especially in research studies in both this country and Europe (e.g., in Romania, Ghioca, 2009). It was used as the basis of an extensive evaluation of summer moisture availability in North America by van der

Schrier et al. (2006). Dai (2011) used both the PDSI and the scPDSI in a global study of moisture conditions, and noted that while the scPSDI considerably reduced the variation in range of the PDSI, it still exhibits variation in range with geographical position.

Alley (1984) describes Palmer's method for identifying the start and end of a wet spell or dry spell to be "rather arbitrary," and notes that the PDSI exhibits unrealistic variations from one month to the next. This is apparent in the examples from Texas shown in Figs. 16-23. As noted above, the protocol for determining the end of a drought is a prominent feature of the index that, among other things, alters its time response characteristics. Alley (1984) considers the use of thresholds to activate and de-activate terms in the water budget to be unrealistic. The runoff term is unrealistic, as well, as it assumes the entirety of the water excess to run off in one month, whereas a considerable lag would be more realistic. There is no accommodation for frozen precipitation (though this is not a limitation for a Texas application).

Much of the criticism in the literature has focused on the soil water budget. Alley (1984) noted it to be rudimentary, and the designation of 1 inch (2.5 cm) as the water capacity for the upper layer was considered to be "rather arbitrary." Vicente-Serrano et al. (2011) dispute the notion that the PDSI is "physically based" because it incorporates a soil-water budget, and extensively criticize the weaknesses of this aspect of the index, mainly on the basis of the complexity of real soils (texture, depth, chemistry, horizons, vegetation, etc.) compared to the treatment in the PDSI as a two-compartment bucket. They also note that data on field capacity is poor, so there is considerable uncertainty in specifying even this simple input.

Akinremi et al. (1996) applied the PDI to the Canadian prairies and found it to be "inappropriate." They made two modifications. First, the Z-index given by (24) was replaced by $Z_i = (P_i - \hat{P}_i)/SD_j$ in which SD_j is the standard deviation of monthly precipitation for month j. This is analogous to the standardized distribution, such as SPI. This eliminated the need for the empirical K-factors of Palmer. The same technique as that of Palmer was used to scale the resulting index to -4 for extreme drought. Second, the soil-moisture part of the Palmer moisture was replaced with a more advanced six-layered soil-moisture budget. On the other hand, Dai et al. (2004) and Dai (2011) used a global data set of PDSI values and found the PDSI to be well correlated with measured warm-season soil moisture.

The PDSI has been compared to other indices. Oladipo (1985) compared the PDSI to the rainfall anomaly index (RAI) and the Bhalme and Mooley index (BMI), see Section 3.1.2 above. The time series of the three proved to be quite similar, RAI and BMI being nearly identical. Oladipo found the Palmer index to respond somewhat more sluggishly to the development of drought but more quickly to pluvials than the other two indices.

Most comparisons have been made to the SPI. McKee et al. (1995) compared the SPI to the PDI over a range of time scales for the former, and found the maximum correlation between the two (about 90%) to occur at the time scale of 12 months. Mavromatis (2007) compared the original PDSI to the scPDSI in Greece, and found that the relative performance between the two was close, depending upon the region of the country. Both indices outperformed the SPI. Lloyd-Hughes and Saunders (2002) found the PDSI and the SPI to be "strongly correlated at all time scales" over Europe, but there are differences in duration and frequency of droughts. Except for spatial distribution of drought incidence, they found the SPI-12 to be "nearly identical" to the PDSI. Corollarily, they found variability in the PDSI to originate mainly in the precipitation input, consistent with the results of Guttman (1991). Dai (2011) in a study of global drought found the PDSI to be superior to both the SPI and the SPEI (Section 3.1.4.3).

Guttman (1998) performed a spectral analysis of a long-time record of both the PDI (technically, the modified PDSI) and the SPI (technically, using a growth-curve cumulative distribution for the rainfall data). He found a characteristic shape of the spectrum of SPI with regional consistency across the continental U.S., but a variety of spectral shapes for the PDI that were regionally inconsistent. His results (which to this reviewer seem strange) evidence minima in coherence between the two indices at ½M and M, the time scale of the SPI, but approaching 90% at longer periods.

Mo and Chelliah (2006) created a "modified" Palmer drought severity index (which is not related to the PMDI product of NCDC), in which the soil-water terms are replaced by data from the 1979-2004 North American Regional Reanalysis, a hybrid data-analysis and atmospheric circulation model. Despite the obviation of many of the assumptions used by Palmer, including the empirical *K* factor, the results proved to be "overall … fairly similar to the PDSI." Another related index is the Surface Water Supply Index (SWSI) proposed by Shafer and Dezman (1982). This is primarily suited to regions with substantial precipitation in solid form that is retained on the watershed until the spring melt. Because this circumstance has little relevance to Texas, the SWSI offers little additional utility beyond that of the PDSI itself, and is therefore not further considered in this review.

3.1.4 Precipitation-evaporation indices

The notion that aridity can be measured by the degree to which precipitation P fails to satisfy the demand for water, *viz*. evaporation E or potential evapotranspiration PE, dates back at least a century, as represented by the definition of "precipitation effectiveness" of Transeau (1905) as the ratio of annual precipitation to annual evaporation. Thornthwaite (1931) developed a regression of P/E on precipitation and temperature, allowing the computation of this ratio from these data alone, from which he constructed maps of the U.S. of "precipitation effectiveness." (Actually, his index was the annual sum of monthly values of P/E, based on the period-of-record mean precipitation and temperature for each month. His complete classification of climate required a similarly derived index of thermal effectiveness.)

The Palmer indices are not included in this category even though potential evaporation is an important component. This is due to the Palmer index being dominated by rainfall, perhaps because when $P - PE < 0$, in effect PE is removed from the calculation, see equation (15) above. Retaining PE in the depiction of precipitation deficiency even though actual evapotranspiration is physically zero is considered by many workers to better represent the shortfall below vegetation requirements and therefore the impact of drought conditions (e.g., Vicente-Serrano et al., 2011).

The simplest such index is the difference $P - PE$ itself. This has been employed in Canada for drought studies (e.g., Hogg, 1994; Sauchyn et al., 2002; Marchildon et al., 2007), where workers find the fact that it is a physical quantity, i.e., the actual water deficit, to be useful. There seems to be a close association between values of this index and the boundaries of forest, parkland and grasslands in central Canada.

3.1.4.1 Aridity Index

By the time of publication of Thornthwaite (1948), his thinking had evolved to recognize that, first, both the physical process of evaporation and the biological process of transpiration needed to be quantified, and, second, it is not the actual evapotranspiration, but the potential evapotranspiration that measures the demand for water. Thornthwaite proposed a moisture index given by ($P/PE - 1$), in which PE is computed from equation (18). (Upon further rumination, Thornthwaite modified this to weight a moisture surplus greater than a moisture deficit by a ratio of three to two, to account for water storage in soil, which leads to an asymmetry in his humidity/aridity classes. His reasoning is obscure, as it seems to disregard the drought condition in which soil moisture would have been lost and has to be made up by infiltration of surface water.) Roudier and Mahe (2010) used a similar index, which they call the climatic moisture index (CMI), in their study of drought on the Bani River in Mali.

Similar indices were used in the mapping of arid regions by the United Nations Educational, Scientific and Cultural Organization (UNESCO), see Meigs (1952) and UNESCO (1979), and by the United Nations Environmental Programme (UNEP), see especially Middleton and Thomas (1992). The UNEP effort defined the aridity index to be P/PE, using annual means of P and PE, the latter being estimated by the Thornthwaite method, equation (18). A moisture deficit is implied by $P/PE < 1.0$, but UNEP defines a “dryland” by a lower ratio of P/PE. This index was used in a global mapping of drylands, in which the classifications of Table 11 were applied, and has been employed to track increased desertification across the earth.

Table 11
Climatic dryland moisture zones of UNEP, from Middleton and Thomas (1992)

designation	*P/PE range**	*fraction (%) of earth land surface*
Hyperarid	[0.00, 0.05)	7.5
Arid	[0.05, 0.20)	12.1
Semiarid	[0.20, 0.50)	17.7
Dry subhumid	[0.50, 0.65)	9.9
Humid		39.2
Cold		13.6

* [*a*,*b*) designates a numerical interval inclusive on the left, that is, all values x such that $a \leq x < b$.

3.1.4.2 Reconnaissance drought index (RDI)

The reconnaissance drought index (RDI) was devised by Tsakiris et al. (2007, see also the earlier publication Tsakiris and Vangelis, 2005) and has been mainly applied in the Mediterranean countries. The basic metric of the RDI is the ratio of precipitation to potential evapotranspiration

$$\alpha_i = \frac{P_i}{PE_i} \tag{27}$$

for the ith month in the time series, or annually

$$\alpha_k = \frac{\sum_{j=1}^{12} P_{jk}}{\sum_{j=1}^{12} PE_{jk}} \tag{28}$$

where the time series is arranged as K rows x 12 columns, K denoting the number of years in the time series (and should not be confused with the K factor in the Palmer index). Here j=1 can correspond to any month in the calendar, not necessarily January. Most applications of the RDI use the water year, so that j=1 corresponds to October. (Note that the annual α is not simply an average of the monthly values.) For each year k, a normalized index is computed

$$\mathrm{RDI}_k = \frac{\alpha_k}{\overline{\alpha}} - 1 \qquad (29)$$

in which $\overline{\alpha}$ is the average over the K years in the record. The RDI is defined to be the standardized normal variate obtained from the data set $\log\{\alpha_k\}$, $k = 1, \ldots, K$, with mean $\overline{\log\{\alpha\}}$ and standard deviation σ,

$$\mathrm{RDI}_k = \frac{\log\{\alpha_k\} - \overline{\log\{\alpha\}}}{\sigma} \qquad (30)$$

Here, as before, log denotes the naperian logarithm. Tsakiris et al. suggest the same thresholds as the SPI (Fig. 10).

In two case studies in Greece, Tsakiris et al. (2007) compared the performance of the RDI with the SPI. For the Mornos Basin, the correlation of the RDI with the SPI was found to be 0.98, and with the decile method 0.90, and for the Nestos Basin these respective correlations were 0.90 and 0.87. Despite these similarities, these authors conclude that the RDI "is expected to be a more sensitive and more comprehensive index." A similar investigation in Cyprus is reported by Pashiardis and Michaelides (2008), who found the correlation to be about 0.95, and concluded that either index was serviceable, the 12-month index better exhibiting the "main" hydrological droughts and wet/dry periods.

The documentation of the details of computation of the RDI is lacking. The annual index α_k is equivalent to the aridity index of the FAO (cf., e.g., George et al., 2010; Nachtergaele et al., 2011). The index α_k can be computed for a period k less than 12 months, notably for 3, 6, and 9 month analyses. The exact procedure is designed for operational use specific to Greece, in which the water year is the basic reference period, and $j = 1$ corresponds to October. An identical procedure is employed with the Streamflow Drought Index (see Section 3.2.1), in which the operational usefulness is more transparent. The normalized index (29) has been little used, and seems to have fallen by the wayside in recent years.

In an application to Iran, Zarch et al. (2011) evaluated the RDI, but aggregating the α data for various averaging windows as is done with the SPI. Their method generally followed that of Tsakiris et al. except instead of a log-normal distribution to fit the data set α_k, which leads to (30), a gamma function was used, analogous to the SPI, see 3.1.2.1 above. Separately, the SPI was computed as well, and over forty meteorological stations, the correlation coefficient between the two indices was very high, averaging 0.94 for the 3-month time scale, declining to 0.84 for 24-month.

3.1.4.3 Standardized precipitation-evapotranspiration index

One problem with all of the above incarnations of an index comparing P and PE is that the index is formed from a ratio, such as P/PE. The ratio tends to distort the effects of PE through increased sensitivity to small values and dampening of variability for large values. In the physics of the water budget, it is the difference $P - PE$, not their ratio, that is operative, so one might anticipate that an index incorporating this difference, but retaining the simplicity of the RDI might be more suitable.

Such an index has recently been proposed by Vicente-Serrano et al. (2010b), called the standardized precipitation-evapotranspiration index (SPEI). A time series of PE_i is constructed from the monthly time series of temperature and precipitation using the Thornthwaite (1948) method, equation (18). From this, a monthly time series of the difference $D_i = (P_i - PE_i)$ is calculated. In the same manner as the SPI, a "time scale" of M months is selected (see Section 3.1.2.1), and this governs the extraction of subsets of sequential values of D_i as given by equation (6). The accumulation of each subset is then determined

$$S_m = \sum_{i=m-M+1}^{m} (P_i - PE_i) \tag{31}$$

analogous to (7), where $m = M, M+1, \ldots, N$, and as before the same symbol is used for the subset of a sequence of S_m and its cumulative. Using data from eleven stations around the world (including Albuquerque and Tampa in the U.S.), Vicente-Serrano et al. (2010b) examined several functions for fitting the cumulative frequency distribution of the time series S_m, including the gamma, Pearson III, and lognormal. Little difference was found among these and the log-logistic, but these workers favored the log-logistic because of its slightly better fit to extreme values. With $L(S_m)$ denoting the ogive of the log-logistic distribution of the M-month cumulatives, the SPEI is equal to $\mathscr{N}^{-1}\{L(S_m)\}$ where as before $\mathscr{N}$ is the ogive of the standardized Gaussian distribution. Vicente-Serrano et al. (2010b) recommend the same moisture categories as the SPI, see Figure 10. Vicente-Serrano et al. (2010a) present a 0.5° gridded data set of SPEI values calculated on time scales ranging 1- to 48-months. They compare the SPEI to the PDSI, finding that the latter is well-correlated with the SPEI for time scales ranging 12- to 18-months, but not at shorter time scales.

A similar index was proposed by Ellis et al. (2010) for use in the Colorado basin (of the West). Like the SPEI, this index computes a monthly time series of the difference ($P_i - PE_i$), and aggregates the monthly data into windows of 3-, 6-, 12-, 24-months, etc., exactly like the SPI. However, instead of fitting a probability distribution then back-transforming to the standard Gaussian, each set of S_m is ranked as an empirical cumulative distribution, and the deciles, expressed as percentiles, become the index. Ellis et al. name this the hydroclimatic index (HI). Because the index is intended to track drought conditions, its threshold values are defined only for the lower frequencies of the ogive, as shown in Table 12, but it clearly could be extended to include pluvials.

The SPEI would appear to combine the best traits of the SPI, the RDI (and related indexes) and the PDSI, because it is a simple index requiring only time series of temperature and precipitation for its calculation, has the built-in ability to select a time-scale of accumulation, i.e., a time filter, incorporates both rainfall and potential evapotranspiration into the index, and does so in a manner more consistent with the governing physics. The most complicated feature of the index

Table 12
Moisture condition classes for Hydroclimatic Index (HI), see Ellis et al. (2010)

designation	*HI value (%)**
abnormal dryness	(25, 40]
moderate drought	(15, 25]
severe drought	(5, 15]
extreme drought	≤ 5

* $(a,b]$ designates a numerical interval inclusive on the right, that is, all values x such that $a < x \leq b$.

is establishing the non-Gaussian probability distribution fit to the basic time series indicator, equation (31), and the transform of this distribution to the standardized Gaussian. Once the distribution is selected and the parameters developed for its application, the calculation can be set up and easily applied thereafter. The HI of Ellis et al. (2010), above, is even simpler to evaluate because it proceeds directly to the empirical ogive. As this report was nearing completion, a paper by Vicente-Serrano et al. (2012b) appeared, comparing global values of the SPEI to the SPI and PDSI, and recommending the SPEI as superior to these alternatives.

3.2 Streamflow-based indices

For Texas, streamflow is perhaps the most prominent indicator of moisture conditions, because it measures the water that is available in riparian corridors and for storage in the state's reservoirs. Streamflow is an integrated response to precipitation and therefore has memory. Unlike the precipitation indices of the previous section, whose memory is dictated by the objectives of the user or the intuition of the formulator of the index, streamflow represents the physical reality of the transformation by the watershed of precipitation into concentrated, accessible water. Granting this, streamflow exhibits pronounced excursions in response to storm events, whose time details are irrelevant to the present concerns but whose incremental volumes to the stream are central. Some preprocessing of streamflow data is therefore necessary to expose the longer-term effects of storms on the landscape. As noted earlier, the smallest time resolution employed here is the month.

Analogous to the precipitation time series of Section 3.1.1 above, one of the simplest streamflow indicators is the monthly time series of streamflow itself coupled with a threshold value. Again (Section 3.1.1), the theory of runs can be employed to address the statistics of periods of above- or below-threshold sequences of streamflow. As with runs in precipitation series, the creation of short pluvials or droughts that interrupt longer, more significant events by breaks in the runs can be surmounted by pre-processing the time series data. Both a moving average and a cumulant (run-sum) strategy have been used with streamflow, and a few variations on these are addressed below. Hisdal et al. (2000) provide an overview of this approach. Tallaksen et al. (1997) present a useful treatment of the threshold approach for streamflow with case studies of two Danish watersheds.

3.2.1 Standardized runoff index

As is the case for precipitation, it is natural to define a streamflow index in terms of the ratio of departure from the mean (or some other suitable reference) to the standard deviation (see Section 3.1.2). There are several examples in the literature, e.g. Beran and Rodier (1985), Ben-Zvi

(1987), and Zaidman et al. (2001), though not dating as far back as similar precipitation indices, perhaps because the period of record for streamflow is generally much shorter than rainfall. Karavokyris and Kaimaki (2010) proposed a "hydrologic year runoff index" for use in Cypress given by $(Q - \overline{Q})/\sigma$ in which Q is annual streamflow, $\overline{Q}$ the historical average annual streamflow and σ the standard deviation of the annual streamflows.

The success of the SPI has motivated a similar index for streamflow, in which monthly flows are accumulated in a window of specified length, using equations (6) and (7) in Section 3.1.2, above. The cumulants are fitted by an appropriate ogive function and transformed to a standardized Gaussian. Shukla and Wood (2008) employed such an index, which they called a standardized runoff index (SRI), together with a watershed-runoff model (rather than measured streamflow data). They found the two-parameter log-normal distribution to perform better than the gamma at higher flows, though the gamma appeared marginally better at low flows. Wen et al. (2011) used what they called a standardized flow index (SFI), defined identically to the SPI, to evaluate the hydrology of the Murrumbidgee River, NSW, Australia, downstream from regulation by releases from a hydroelectric dam. The gamma distribution was applied to convert flows to a Gaussian variate. They determined that hydrological droughts as measured by the SFI were less severe than meteorological droughts as measured by the SPI. Vicente-Serrano (2012a) apply to the Ebro basin what appears to be the same index, which they call the standardized streamflow index (SSI, though in the title to their paper, they name it the streamflow drought index).

A new index has been proposed by Nalbantis and Tsakiris (2009, published on-line in 2008), see also Nalbantis (2008), which they call the streamflow drought index (SDI). This was developed for application in Greece, and is based on a strategy of accumulation specific to a mediterranean hydroclimate, with maximum rainfall in winter and spring and minimum in summer. The water year (from October of the previous year through September) is therefore a fundamental period of time. The time series of monthly flows Q_i is arranged as Q_{jk} for $j = 1, 2, \ldots, 12$, and $k = 1, 2, \ldots,$ K, that is, K rows x 12 columns, $j = 1$ corresponding to October, so that each row is a water year, and K denoting the number of water years in the time series. For each water year k, four "reference periods" of data are used, ending respectively 31 December (3-month), 31 March (6-

month), 30 June (9-month) and 30 September (12-month). These are accumulated for each water year k as:

$$R_{k\,n} = \sum_{j=1}^{3n} Q_{jk} \quad \text{for } n = 1, 2, 3, 4 \tag{32}$$

Nalbantis and Tsakiris (2009) desire to standardize this, but recognizing that the $R_{k\,n}$ data, like rainfall, may be skewed positive, they first normalize their data by a transform, for which they select the log-normal. The streamflow drought index (SDI) is then defined as

$$SDI_{k\,n} = \left(\log\{R_{k\,n}\} - \overline{\log\{R\}}_{\,n}\right) / \sigma_n \tag{33}$$

where $\overline{\log\{R\}}_{\,n}$ and σ_n denote the mean and standard deviation* of the naperian logarithms of the $3n$-month sums over the K water years in the data set. The same thresholds as the SPI, Table 7 and Fig. 10, are used except that 0.0 ≥ SDI ≥ -1.0 is classified as "mild drought". However, these workers found it desirable to re-define the drought categories ("states") into numeric categories that appear closer to the PDSI, see Table 13.

For a test of the methodology, Nalbantis and Tsakiris (2009) computed both the SPI and SDI for the Evinos basin in central Greece. The 6-month and 12-month values were found to be nearly identical for both indices, which is an expression of the strong Mediterranean seasonality of rainfall — a wet season of winter and spring, and a summer dry season. In general, the two indices were highly correlated. The regressions of SDI on SPI exhibited slopes less than one, which these authors interpret as a meteorological drought producing a hydrological drought of less severity, mainly due to lags in runoff. They state that the main utility of the regression is to estimate SDI based on SPI, because streamflow is generally not available in real time. This may be true of Greece, but it is not true for the United States. It also raises the question of why employ the SDI at all, rather than simply use the SPI.

* See note, page 48.

Table 13
Drought categories for use with the Standardized Drought Index (SDI), from Nalbantis and Tsakiris (2009)

Category	*Description*	*Criterion*	*Cumulative probability (%)*
0	non-drought	SDI ≥ 0	50.0
1	mild drought	[-1.0, 0.0)	34.1
2	moderate drought	[-1.5, -1.0)	9.2
3	severe drought	[-2.0, -1.5)	4.4
4	extreme drought	SDI < -2.0	2.3

Recently, Vicente-Serrano et al. (2012a) provided a detailed study of the appropriate distribution function for use with the SRI, SFI, or SDI, based upon monthly streamflows from the Ebro in Spain. The best distribution varied with position along the river, which these workers interpret as being due to physical features of the channel and watershed. They also found the best distribution to vary with season, apparently due to the synoptic features that drive rainfall. The high flashy flows of spring and summer are best represented by the generalized Pareto distribution, but the frontal rainfall of fall and winter have a smaller range of intensity for which the log-logistic and lognormal are preferable. This notwithstanding, these authors found that little differentiation in the final values of the index could be made between the different probability distributions and the strategies for fitting the distribution. Some of these distributions were judged unsuitable because the computed mean and standard deviation were not sufficiently close to the theoretical values of 0 and 1, resp. Their recommendation was either the generalized extreme value or the log-logistic.

3.2.2 Mass curve index

A streamflow-based method particularly useful for diagnosis and display of the occurrence of surfeit and drought periods is provided by the so-called residual mass curve (Rippl, 1883, McMahon and Mein, 1986, see also Ward and Proesmans, 1996), given by the cumulative sum:

$$\sum (Q - \overline{Q})$$

where $\overline{Q}$ is the period-of-record mean. A period of below-average flow is exhibited in the time plot of the cumulative-residual-flow as a declining trend in the curve. Similarly, a period of above-average flow is a rising trend in the curve. For hydroclimatological purposes, the "period" in this statement must be defined as some minimum duration. The trends can be then be quantified as the least-squares line passing through the streamflow data for each period so defined. In principle, this can be computed automatically by stepping through the data file, computing a trend line for the period specified (with a little more logic to handle the case of durations longer than the minimum period). However, it is simple to carry this out manually by inspecting a time plot of the streamflow series. This was recently used in analysis of the hydrology of the Klamath River in northern California (Ward and Armstrong, 2006). An example from this application is shown in Figure 24.

More quantitatively and more generally, a drought can be defined to be a period of at least N months whose mean flow is less than some fraction $f \leq 1$ of the period-of-record average flow. Similarly, a surfeit period (or simply "surfeit") or pluvial can be defined as a period of at least N months whose mean flow is greater than some factor $g \geq 1$ of the period-of-record average. Droughts are diagnosed when the general downward segment of the streamflow time series is steeper, for this minimum length of N data points, than the straight line

$$R(t) = \sum(Q_o - \overline{Q}) + (f - 1)\ \overline{Q}\ (t - t_o) \qquad (34)$$

where $(t_o, \sum(Q_o - \overline{Q}))$ is the first point of the declining segment, and the slope of the line is

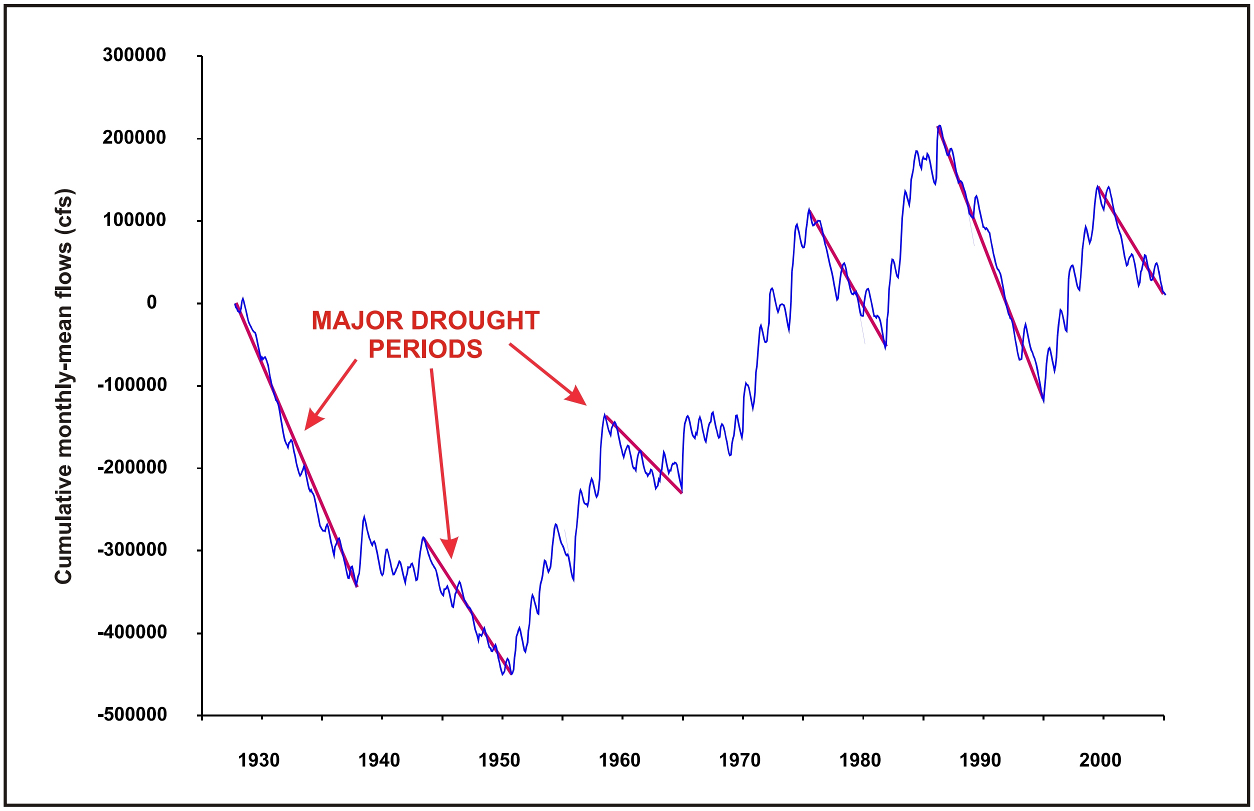

Figure 24 - Cumulative monthly flow diagram, Klamath at Orleans (vertical units arbitrary), from Ward and Armstrong (2006)

$(f - 1)\,\overline{Q}$. Since $f \leq 1$ the slope of the line (34) is ≤ 0. An analogous method is used for a surfeit period, in which the factor $(g - 1)$ replaces $(f - 1)$.

Two aspects of quantifying a surfeit/drought period are its intensity and its severity. The mass-curve method is particularly useful in this regard. Intensity can be defined as the degree by which the flow falls above/below the diagnostic criterion $g\,\overline{Q}$ for surfeit and $f\,\overline{Q}$ for dought. The main utility of this approach is that it provides a convenient, intuitive criterion for terminating the drought/pluvial period. The duration is defined by the time from the first point of the rising or declining segment to the intersection with the mass curve of the line (34) for drought or its analog for surfeit. The intensity of the surfeit/drought is measured by the slope of the regression line through the flow data for the surfeit or drought period. The steeper the slope of the regression line, the more intense the surfeit or drought in terms of average flow surplus or deficit. Severity is the product of duration and intensity, measured by the maximum deficit below the criterion line (34) attained by the least-squares trend line (not the data of the time series), and likewise for the surfeit as the maximum surplus above its criterion line. Alternatively, severity can be computed as duration times the slope of the regression line.

Those familiar with the use of the residual mass-curve methods for reservoir capacity estimation will recognize the deficit volume as the capacity necessary for a theoretical reservoir to provide a firm yield of the criterion slope, in this case, the fraction f of the period of record mean flow. The parameter f basically specifies the reference flow, as a fraction of the period of record mean flow $\overline{Q}$, relative to which a surfeit/drought condition is defined. In the case of the Klamath, the period-of-record mean itself was taken to be the reference flow (i.e., $f = 1$), so the slope of (34) is zero, and a minimum drought period (i.e., "major") was defined to be a period of at least seven (7) years. Clearly, the hydroclimatology of the Klamath is different from a Texas river. These major droughts are indicated as red trend lines in Fig. 24.

This residual mass curve (RMC) method was suggested to the Guadalupe-San Antonio BBEST during its work on formulating flow standards for the basin (which overlapped with the period of study of the present project). Interest was, of course, focused on drought events. In this case, a drought was defined to be referenced to 60% of the period-of-record mean flow (i.e., $f = 0.6$) and sustained for a period of somewhat more than a year. The RMC time plot is shown in Figure 25, based on the 1942-2009 record of monthly flows into San Antonio Bay (the combination of gauged flows in the San Antonio and Guadalupe Rivers and estimates of nongauged runoff), see Ward (2010). Equation (34) was applied to define drought periods. Ten drought periods were found, numbered from 1 to 10 in Fig. 25, of which 1 is the notorious Drought of the 50's, and 2 is the drought of the 60's, see Table 14. TWDB has implemented this method in an EXCEL® workbook.

At present, an objective method for specifying f is unavailable, but rather must be based on intuition founded on familiarity with the river basin under consideration. (For a water-rich

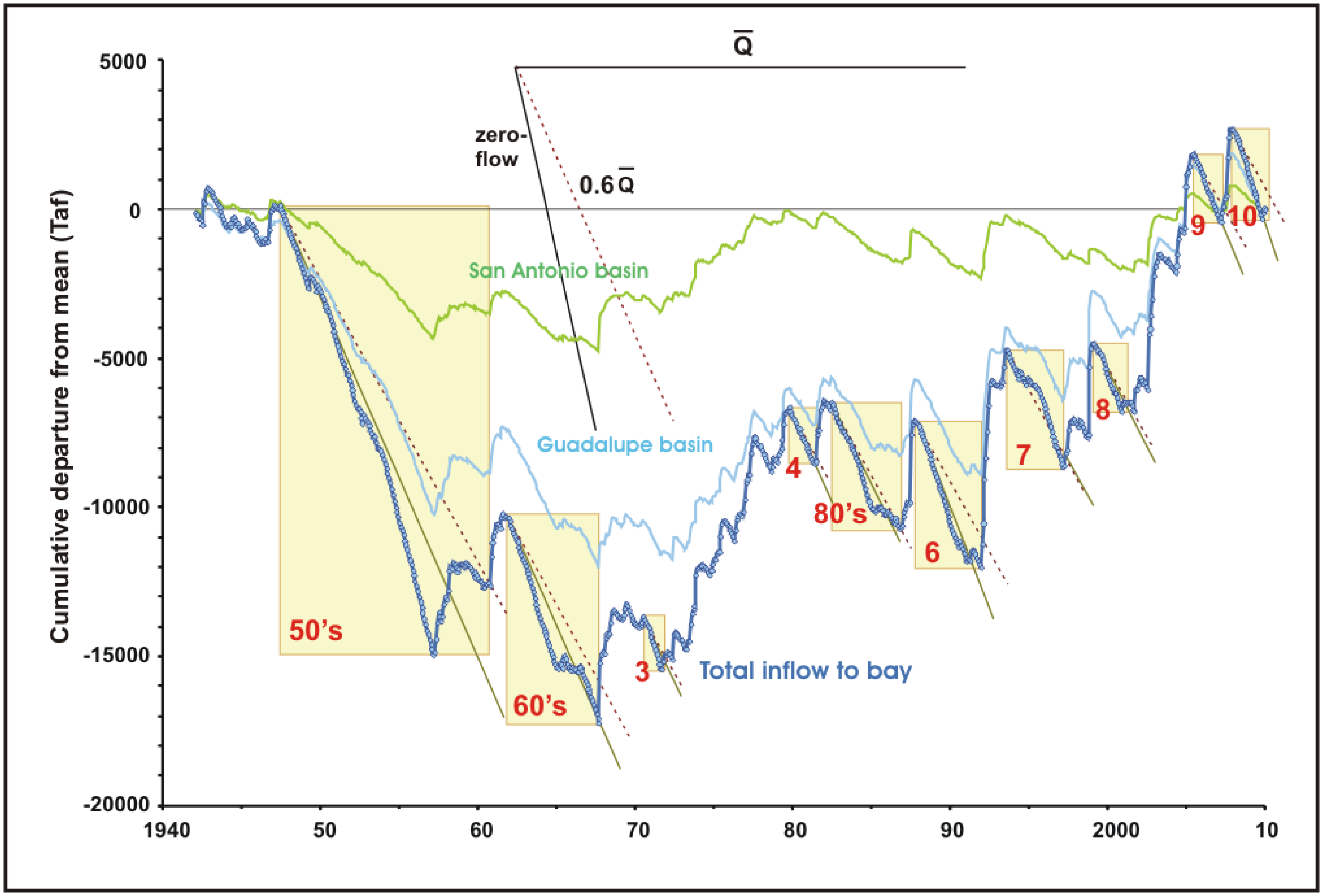

Figure 25 - Residual mass curve of monthly inflow time series of San Antonio Bay, prominent droughts indicated by constant-flow (0.6 $\overline{Q}$) criterion lines and regression lines, from Ward (2010)

Table 14
Droughts during 1942-09 period for flows into San Antonio Bay, based upon mean flow < 0.6 $\bar{Q}$, from Ward (2010)

Drought ID	*drought period**		*best-fit slope*	*max depletion*	
	start (year)	*end (year)*	*(Taf/yr)*	*volume (Taf)*	*date (year)*
50's	47.50	60.75	-1209	5777	57.17
60's	62.00	68.00	-1182	2065	65.00
3	70.58	72.17	-1089	716	71.58
4	79.83	81.42	-1173	378	81.25
80's	82.50	86.92	-1100	1117	85.00
6	88.00	92.00	-1374	1697	91.00
7	93.67	97.25	-933	509	97.17
8	99.25	01.62	-1025	632	00.83
9	05.58	07.33	-1280	661	07.00
10	08.00	10.00	-1508	1240	09.75

* Dates are given as years after 31 Dec 1900 omitting the hundreds unit. 05.58 is therefore 0.58 of a year into 2005, i.e. day 0.58 x 365 of 2005.

region like the Klamath, the mean flow itself is clearly a viable candidate. For Texas, such a choice would yield 50-year droughts for some rivers.) Additional research on how best to specify *f* is needed. Alternatively, a better formulation for objectively determining the termination points of a drought needs to be devised.

3.3 Reservoir-based indices

The TWDB staff is intimately familiar with reservoir operation and data sources, so there is little to be added in this section. Generally, stage of a reservoir is routinely logged at some temporal resolution, typically daily, by the operator of the reservoir, and converted to contents using the capacity-stage relation. This relation is prepared by the designer of the reservoir based on pre-inundation topography, and typically updated by post-construction bathymetric surveys. In Texas, major reservoirs are operated by the Corps of Engineers, the Bureau of Reclamation, river authorities, cities and municipalities, industries, and power companies. The TWDB presents a useful compilation of percent of capacity data for most of the state's water-supply reservoirs on its website, including graphic displays.

In some respects, the contents of water in a reservoir can be viewed as the most basic metric for water management. For a supply reservoir, the contents relative to probable inflow measures the vulnerability of the water supply, and in a drought condition the contents represents the immediate availability of water to meet demands. The analogy of managing expenses by the balance of a bank account — and the need to ration expenses during hard times — comes immediately to mind. For a flood-control reservoir, the contents relative to probable high inflow likewise measures vulnerability (but the bank account analogy fails because most of us would not regard a sudden large influx of income to be problematic).

Design and operation of flood control reservoirs comprise a well-established specialty based on the characteristics of storm-runoff events in the specific river, and is therefore not included in this study. For water supply, in principle the allocation of water according to firm yield of a supply reservoir is a zero-failure strategy, so there should be no need to further consider this topic either, and this chapter should now end. But the fact is that with burgeoning water demands, increasing appropriations to junior and "interruptible" permits, and recurring droughts with the specter of sharply diminished reservoir stages, the need for a more reactive water management policy has arisen in several of Texas basins and may become necessary in others.

3.3.1 Reservoir water budget

Let $R(t)$ denote the contents of a reservoir of capacity V, which could be the volume below the spillway of a water-supply reservoir, or the water-supply allocation in a reservoir that also has a flood-control pool. The net inflow to the reservoir is given by $Q(t) = Q_s(t) + P(t) – E(t) + Q_g(t)$, where Q_s is streamflow (perhaps from multiple tributaries), P and E are the flows across the surface of rainfall and evaporation, resp., each equal to the surface flux rates multiplied by the reservoir surface area, and Q_g is the influx through the bed of the reservoir, typically negative in Texas. In addition there is a withdrawal rate from the reservoir given by Q_b which is tracked separately. The volume stored in the reservoir is then given by:

$$R(t) = \max\left\{ \min\left\{ [R_i + \int_{t_i}^{t} (Q - Q_b)\ dt], V\right\}, 0\right\} \tag{35}$$

where t_i = sequence of times bounding periods of extrema where $R(t) = V$ or $R(t) = 0$

$R_i \equiv R(t_i)$ where $t_i \le t \le t_{i+1}$ for $i = 1, \ldots, n-1$

$t_o \equiv 0$

$R_o \equiv R(0)$

The sequence of times t_i is determined from the behavior of the integral $\int_0^t (Q - Q_b)\ dt$. The time period of the integral may be divided into intervals in which the integrand $(Q - Q_b)$ is negative or nonnegative. Within each interval in which $(Q - Q_b) \ge 0$, $R(t)$ may acquire the value V. If it does so, then the first point at which $R(t) = V$ and the starting point of the next interval in which $(Q - Q_b) < 0$ define an interval (t_j, t_{j+1}) in which $R(t) = V$. This is called a full interval and the bounding times t_j and t_{j+1} are full points. In the engineering literature, this is referred to as a spill or wastage interval, because the contents in excess of V are "spilled" downstream. Within each interval in which $(Q - Q_b) < 0$, $R(t)$ may fall to zero. If so, then the first point at which $R(t) = 0$ and the starting point of the next interval in which $(Q - Q_b) \ge 0$ define an interval (t_k, t_{k+1}) in which $R(t) = 0$, called a fail interval, because the "demand" Q_b is not met, with bounding times the fail points t_k and t_{k+1}. The sequence of times t_i in (35) above is given by

$$\{ t_i \mid t_i \text{ bounds period(s) of extrema} \} = \{ t_j, t_{j+1} \mid \text{full points} \} \cup \{ t_k, t_{k+1} \mid \text{fail points} \}$$

after the union on the right-hand side has been ordered chronologically. The initial time t_o may or may not be a fail point or a full point, depending upon the value of R_o. In practical application, the integral is discretized into a sum over increments of time, typically one month.

Ward and Proesmans (1996) used this framework to organize and review the theory of storage and yield of water-supply reservoirs, from which (35) is taken (with some minor rearrangement of terms and correction of an obvious typographical error). Many of the manual reservoir design methods in the literature can be reduced to a manipulation of (35) (see Ward and Proesmans, 1996). Sometimes it is convenient to collect $P - E + Q_g - Q_b$ into a single term, or simply neglect $P - E + Q_g$, so that $Q(t) = Q_s(t)$ is only streamflow, a nonnegative time series. Sometimes a constant flow, usually the period-of-record mean $\overline{Q}$, is subtracted from each term in (1) to make arithmetic or graphing easier to handle, such as in the "residual mass curve" method discussed in 3.2.2 above.

This is a mathematical model of a simplified reservoir. There is no provision for a controlled release or spillway rating, for peripheral watershed inflows, or operating rules or drought triggers, or for preserving flows for senior rights up- and down-stream, and many of the complexities of a real reservoir are hidden in the mass-budget terms, for example the dependency of $P - E$ on the reservoir stage-capacity relationship. What is useful about this model and important in the present context is the general behavior of the $R(t)$ time series (35). A time interval bounded by a full point $R(t_i) = V$ and a fail point $R(t_{i+1}) = 0$ is a depletion or *drawdown* period. An interval bounded by a fail point and a full point is a filling or *storage* period. The draft Q_b that eliminates all fail intervals so that there remain only discrete fail points is the *firm yield* Q_f. Theoretically, there can be multiple fail points under a firm yield draft, but in real-world cases this is exceptional, and for practical application there is exactly one. The depletion period culminating in that fail point is the *critical drawdown period*.

In Texas water management, the critical drawdown period is the drought of record. The firm yield is therefore the largest constant draft Q_f that can be supplied by the reservoir over the period of record of inflow data. It is the draft that drives reservoir contents exactly to an instantaneous zero during the drought of record. So long as the drought of record is the worst possible drought — a monumental assumption — and the draft does not exceed the firm yield, water demands will always be met.

3.3.2 Reservoir contents-based criteria

One does not look far to find an example of reservoir management in which the triggers are specified fractions of storage in a reservoir, as that is precisely how the water supply to Austin from the Highland Lakes is determined. It is important to observe that this strategy is demand restriction, governed by drought stages based on reservoir contents (as a fraction of capacity) in which limitations on levels and types of water use are specified, and implemented by the local water supplier, e.g., the City of Austin. Similar water-supply schemes are used in other water-supply reservoirs, both in Texas, and elsewhere.

This strategy represents a disconnect from the yield-based allocation strategy outlined in the previous section. The water restrictions are designed to incrementally reduce the demand for water as the severity of a drought, as manifested by reservoir level, increases. When the reservoir is full, the use is typically unrestricted, in some cases through interruptible permits, but as dry conditions persist and reservoir level declines, restrictions are imposed. Usually, rains return, the lake level rises, and restrictions are removed. But early in a drawdown period when uses are unrestricted, one never knows whether this is the beginning of a critical drawdown period. Early diversions in excess of the firm yield could accelerate the fail point as the drought worsens, despite an incremental water-restriction strategy.

The literature review for examples in which basin-wide water management is tied to reservoir levels, or cases in which reservoir management is tied to triggers of a hydrometeorological index, proved futile. There are numerous cases of extensive reservoir development, several of which are in basins with hydroclimatologies similar to Texas, but in every case, water management during drought devolved to locally implemented water restrictions. One such case is summarized below, as an example.

3.3.2.1 Case study: Cyprus

The island of Cyprus in the Mediterranean Sea has historically had a serious water-supply problem (Water Development Department, 2003). The climate is mediterranean (i.e., hot summers and mild winters, moderated by proximity to a mediterranean sea) and semi-arid with dry summers and wet winters, similar in many respects to the Edwards Plateau and South divisions in Texas (NCDC Divisions 6 and 9, resp.). Annual rainfall is around 300 – 1100 mm, averaging about 460 mm over the island (1971-2000 normal), with highest rainfalls (about 80% of the island's total) over the Troodos Mountains in the center of the island. The seasonal rainfall is typical of the Mediterranean, unimodal with maxima in December – February and minima in July – September, forced mainly by disturbances in the westerlies. The higher interior rainfall is orographic, due to interaction of the Troodos massif with synoptic storms in the westerlies (Pashiardis and Michaelides, 2008). About 60% of the annual precipitation falls in December – February.

The island is drought-prone. Like Texas, the hydroclimatology of the watershed amplifies the effect of diminished precipitation on runoff (Ward, 2011). During the 1996-2000 drought in Cyprus, precipitation 75% of normal produced runoff 25% of normal. This is almost identical to the runoff response to precipitation under drought conditions in south Texas determined by Ward (2011). The frequency of drought led to the institution of water management in the colonial period of the early twentieth century and was constitutionalized in 1960 with independence (Water Development Department, WDD, 2003). A Master Plan was put in place in 1971 and major water-supply projects were undertaken. Total reservoir capacity was increased from 5 Mm^3 to the present 328 Mm^3 (WDD, 2009). A revised River Basin Management Plan was prepared in 2011, presently under review, to implement Water Framework Directive (WFD) 2000/60/Ek of the European Union (whose deadline is 2015).

Table 15
Mean annual water budget of Cyprus, 1971-2000 normal, Mm^3/yr
see Pashiardis and Michaelides (2008), Tsiourtis et al. (2007), Iglesias et al. (2007a)

	Island water budget		
Input from atmosphere		*Disposition of surface water runoff*	
Precipitation	4250	Streambed storage	140
Evapotranspiration	– 3450	Diversion for irrigation	40
Net P – E*	800	Retention in reservoirs	110
		Flow to sea	230
Disposition of annual water crop			
Groundwater infiltration	280		
Surface runoff	520		
	Under control of Cyprus government		
Input from atmosphere		*Disposition of surface water runoff*	
Precipitation & evapotranspiration	n/a	Streambed storage †	5
Net P – E*	330	Diversion for irrigation †	35
		Retention in reservoirs	110
Disposition of annual water crop		Flow to sea	50
Groundwater infiltration	110		
Surface runoff	200		
Human consumption		*Sources for human water demand*	
Agriculture	170	Groundwater	135
Domestic & industrial	75	Surface water	105
		Wastewater re-use	5
		Desalination**	35

* "Annual water crop," which separates into subsurface storage and runoff.
† Not given in above citations, therefore estimated.
** Recent estimates are about double this value.

An approximate water budget is presented in Table 15. There are apparent conflicts between some of the data in Pashiardis and Michaelides (2008) and those presented in Iglesiades et al. (2007a). These are resolved by recognizing that the hydroclimatology of Pashiardis and Michaelides (2008) represents the entire island, while the data of Iglesiades et al. (2007a) and Tsiourtis et al. (2007) are confined to the area under the control of the Cyprus government. This separation is made in Table 15. The areas *outside* of the control of the government are (1) the regions in which the streams and rivers are small and remote, so that development is expensive, and (2) the northern 37% of the area of Cyprus occupied by Turkey since 1974. Water supply for human uses is made up of groundwater withdrawal, surface water retention, desalination, and a small amount of wastewater re-use. Groundwater extraction exceeds recharge. Iglesiades et al. (2007a) state that the normal groundwater withdrawal of 135 Mm^3/yr exceeds the estimated safe yield of 110 Mm^3/yr. The major aquifers are indeed being depleted at alarming rates.

A slow crisis has developed in the past thirty years from a conflict of trends in the water balance: precipitation has been trending steeply downward, air temperatures upward, and population sharply upward, all of which have conspired to entail more frequent droughts and near failures of water supply. Since 1970, the island suffered six major droughts. In the latest, 2004-2009, the October 2007-September 08 rainfall was *half* normal. The amplifying effect on runoff of the precipitation deficit drove an already stressed water supply to critical, as the reservoirs were drawn down to nearly empty. Drastic rationing was imposed and water had to be shipped to the island.

Hydroclimatological studies have been carried out for Cyprus based on some of the indices reviewed previously. Michaelides and Pashiardis (2008) used the SPI to monitor the drought in 2007-08. Interestingly, a different threshold system for the SPI was adopted for their study, given in Table 16, apparently because better differentiation of the relative intensity of drought conditions was needed. But these studies do not appear to have been directly linked to the problem of water supply. Tsiourtis et al. (2007), for example, found no correlation at all between SPI and water supply quantities. This is attributed to the effect of reservoirs in capturing and detaining streamflow. Further, there was little correlation between SPI and agricultural production, except for crops such as wheat farmed without irrigation (i.e., purely rainfed).

The present water-supply system is complex, consisting of over 100 dams and a complicated network of conveyances (WDD, 2003, 2009). Water management in Cyprus is presently in a state of flux. In 2010, the Water Development Department was empowered to provide integrated management of water resources. Analysis and planning of the WDD encompasses the entire island, though it has limited influence on projects in the north. (More details on the effects of the

Table 16
Moisture condition classes for SPI used in study of Cyprus by Michaelides and Pashiardis (2008), cf. Table 7

designation	SPI *value**
exceptionally wet	≥ 3.0
extremely wet	[2.00, 3.00)
very wet	[1.25, 2.00)
moderately wet	[0.75, 1.25)
near normal	(-0.75, 0.75)
moderately dry	(-1.25, -0.75]
severely dry	(-2.00, -1.25]
extremely dry	(-3.00, -2.00]
exceptionally dry	≤ -3.0

* (*a*,*b*] designates a numerical interval inclusive on the right, that is, all values x such that $a < x \leq b$. [*a*,*b*) denotes an interval inclusive on the left.

Turkish occupation on water management are given by Charalambous et al., 2011.) In the 1960's, the philosophy of "not a drop to the sea" led to the extensive dam and conveyance construction program of the late twentieth century. In the past couple of decades, the concern that water supply be "sustainable," and recognition of the importance of maintenance of downstream flows have led to a more balanced management strategy. This has been accompanied by an awareness of the importance of data in the process, and an associated acceleration of various monitoring programs. Part of the stimulus for this is the European Union WFD, which dictates, among other items, a Programme of Measures (i.e., data collection) and requires a drought management plan (e.g., Estrela and Vargas, 2012).

Pursuant to the EU WFD, a consulting study was commissioned by the WDD to develop a comprehensive drought management plan, reported by Karavokyris and Kaimaki (2010). This plan recommends the SPI-12 (and multiple-year windows) for monitoring drought conditions, supplemented by a runoff index (see Section 3.2.1) evaluated for representative dams to diagnose changes in runoff not detected by the SPI. In addition, for the main water supply systems a dams-storage index (a.k.a. large projects reserves index) is proposed to be used based on the stored volume of each reservoir (or tandem reservoirs) in each system compared to the historical water demands. A schedule of zero to "very significant" cutbacks is tied to the total contents, for each of which a target annual diversion is specified.

An "early warning" index is proposed for establishing "alerts" for an impending drought. This is based on the regularity of the season of maximum rainfall ("wet period") on Cyprus. Starting in December and continuing through April, for each month the runoff volume since October is compared to the historical ogive of runoff at that month for the same calendar period. An alert level is issued if this runoff is less than the 25% occurrence flow, with triggers for "high" and "very high" alerts if the runoff is less than 15% and 5%, resp.

The "pressure" on riverine and riparian ecosystems will be addressed by a monthly regime index comparing the median daily flow in that month to the frequency of occurrence of historic daily flows for the same month. Two triggers are proposed, "important" if the median flow lies within the lowest quartile, and "high" if it lies in the lowest 5%. A "prolonged drought" is defined to be a drought of such rarity "that the conservation of the water bodies' protection measures which are foreseen in the Management Plan will not be achievable," (Karavokyris and Kaimaki, 2010, p. 35), based on a combination of the above indices. Occurrence of a "prolonged drought" will necessitate downgrading of the assigned ecological status.

The Plan includes a tabulation of actions assigned to each alert level inferred from the above indices. The majority of actions involve communication to operators, users and the public. As stated above, the Plan includes tables relating diversions from large projects to the storage status for each, so these diversion are incrementally curtailed. However, these are no allocation specifics as to which users are affected and at what magnitudes. Apparently, these decisions will be left to local operators or water boards. The ecosystem index appears to be used only to establish when it is permissible to curtail releases for environmental purposes.

As the details of this Drought Management Plan have only recently appeared, it is in review and not yet implemented. It may prove useful to Texas to continue to monitor the evolution of the Cyprus situation.

4. DEMONSTRATION CASES IN TEXAS

4.1 Selection of indices and application sites

The desiderata arrayed in Section 1.1.1 at the outset served as criteria for the selection and review of the many moisture-condition indices that appear in the literature, presented in the previous chapter. For purposes of identifying candidate indices for potential utility in Texas, three general criteria were applied. First, values of the index should be routinely and readily available on at least a monthly basis, or easily computable from variables that are available, which combines (ii), (v) and (vi) of Section 1.1.1. Second, there should be extant a long period of record of the index to serve as a foundation for hydroclimatological analyses, which is criterion (vii) of Section 1.1.1. Third, there should be good evidence in the literature of the applicability of the index to the Texas environment, which is a provisional indication that criteria (iii), (ix) and (x) of Section 1.1.1 may be satisfied. In addition to these criteria, qualitative applicability of (i) appropriateness and (iv) communicability, along with the basis of formulation were applied to further winnow the field of choices. These considerations are subjective and border at times on matters of taste. For example, this reviewer admits to a predilection for an index that has some physical basis or motivation, over one whose formulation reeks of artifice. The following indices were selected for additional examination.

The family of Palmer indices (Section 3.1.3) is an obvious choice, because of their long history of application, particularly the PDSI. Further, monthly updates of the PDSI for the state climatic divisions are routinely available from the NCDC website as part of TD-9640. The often erratic behavior of the PDSI due to the re-initialization employed as part of the wet/dry-period protocol is undesirable, however. As an ancillary index, the X-index given by equation (25) may have merit, as it is basically the PDSI without the wet/dry-period protocols. This is readily calculated using the recursive formula (25) from the Z-index (24), which is also a product of TD-9640.

The standardized precipitation index (SPI, Section 3.1.2) has become popular among hydroclimatologists since its introduction. It has recently been recommended for use by the state by Quiring et al. (2007), see also Quiring (2009a, 2009b). This index has been made part of the

NCDC climate division data of TD-9640, so it is also conveniently available and updated monthly. Given that it is usually the longer-term vacillations in hydroclimate of concern in Texas, the SPI-12 and SPI-24 are deemed most suitable as candidate indicators. The SPI-6 is also chosen because occasionally there may be use for an index with short-term responsiveness.

At its core, the SPI is a backward-looking average of some specified duration. The mathematical complexity of the SPI engendered by its conversion to a standardized Gaussian raised the question of whether a conceptually simpler index could exhibit the same responsiveness to moisture conditions. The historical index of the surfeit as a fraction of mean precipitation (Section 3.1.1), equation (4), was selected for this purpose, for which *P* is taken to be a one-year average. Specifically, the suggested index is:

$$N_i = \frac{\overline{P}_i - \hat{P}}{\hat{P}} \tag{36}$$

where $$\overline{P}_i = \sum_{k=i-11}^{i} P_k$$

and $\hat{P}$ is the normal annual precipitation. For this application, this is taken to be the 1971-2000 normal. N_i is a surfeit fraction-of-normal of the 12-month mean ending in the current month *i*. Negative values indicate dry spells and positive values wet spells, and the index is bounded below by -1, so it is skewed positive. Besides this skew, the main difference between this index and the SPI is that N explicitly incorporates a measure of "normalcy" in the index, while the SPI deals with the raw monthly precipitations, leaving the Gaussian standardization to effect "normalization."

The conceptual basis of the standardized precipitation-evapotranspiration index (SPEI, Section 3.1.4.3) would appear to be superior to the SPI because it combines rainfall *P* with a measure of surface water demand by using potential evapotranspiration *PE*, analyzing the difference P – PE. The SPEI, as formulated by Vicente-Serrano et al. (2010b) is not routinely available nor readily calculated, however. Nor has it yet received the validation of widespread use. To explore the potential of this sort of index, the much more facile hydroclimatic index (HI) of Ellis et al.

Table 17
Summary of hydroclimate indices selected for application to Texas watercourses

(a) short-term memory	
SPI-6	HI-6
(b) one-year memory	
SPI-12	surfeit N index
HI-12	
(c) long-term memory	
PDSI	Palmer X index
SPI-24	RMC

(2010) was implemented instead. For comparability to the SPI-12 and the surfeit as fraction of mean N index above, a 12-month window was used. For comparability to the SPI-6, the HI with a 6-month window was also selected.

The residual mass-curve (RMC) index using monthly streamflow (Section 3.2.2) has already found some utility in Texas applications, so is a logical choice. Rather than descending into the unresolved issue of how to identify wet/dry periods using (34), for present purposes the cumulative $\sum (Q - \overline{Q})$ is used, being examined manually for declining trends. Of the other streamflow indices reviewed, the standardized runoff index (SRI), standardized flow index (SFI) and standardized streamflow index (SSI), see Section 3.2.1, are all similar and may offer some potential but at present have not had sufficient application, are not routinely available, and are too complex for ready computation.

In summary, the indices to be tested in Texas are listed in Table 17. Selection of test watercourses was simpler. Gauges were sought on major rivers, exhibiting a range of watershed climatologies, and minimal impacts from upstream reservoirs (though in Texas some upstream reservoirs cannot be avoided on major rivers). The data source for the RMC analysis is the monthly mean streamflows developed from the USGS daily data for each streamflow gauge. The NCDC division-mean values of precipitation, temperature, SPI and PDSI (and its Z-index) are used. The use of a uniform period of data avoids any corruption of the comparative behaviors of the indices due to differing periods of record. Although the periods of record extend back to the early twentieth century for most of the USGS gauges, and back to 1895 for the NCDC data, the period adopted for the test cases is 1945-2011. For a few of the gauges with records that do not extend back to 1945, the RMC is computed on the record that does exist and only compared to the other indices for this shorter record.

The objectives of selective testing on Texas watercourses and of preparing demonstration cases were combined so that each test case for every selected index was also fully developed as a demonstration case. For each selected site, the index time series are displayed graphically. Plots are provided for 20-year intervals, starting in 1950, thereby encompassing the period 1950-2009 (inclusive). The plots are organized by index memory, as grouped in Table 17, to facilitate readability and comparisons. Though no specific discussion is provided in the report sections for the individual gauges, noteworthy behaviors are marked and labeled on the figures. For example, in Fig. 30, at “A” the SPI-12 shows a pronounced pluvial that is not indicated by the other indices. Remarks on the relative behaviors of these indices are given in Section 4.10.

The RMC index is a different sort of parameter compared to the other indices, and must be interpreted differently. First, it is in arbitrary units, since no normalization is performed. However, the range plotted on each figure is a fixed value. This range will vary from gauge to gauge in order to best depict this index. Though the range is fixed, the origin may be shifted up or down so that the full range of variation of the RMC is displayed for each 20-year plot. Second, it is not the magnitude of the index but its slope that is important, more specifically the trend through a series of index data points. A positive slope (or increasing trend) indicates a pluvial, a negative slope (or declining trend) a drought.

For each gauge site, a correlation array is presented between each of the indices, e.g. Table 18. The variables monthly rainfall and streamflow are included with the indices for the correlation calculations (because if they were not, someone would ask for them).

4.2 Sulphur River at Talco

Talco is located near the eastern boundary of the NCDC North Central climatic division (Division 3), and almost all of its 3650 km^2 watershed upstream from this gauge lies in this division. There are no major reservoirs upstream from the gauge site. There is a relatively high density of North American wood ape (a.k.a., bigfoot) sightings in this basin, mainly in the Sulphur bottoms but some in the Talco watershed. Unfortunately, the record on this gauge is shorter than desirable, starting in 1956. In Figs. 32-34, the range of the ordinate for RMC is 65K, and the origin varies from figure to figure. Table 18 summarizes the pairwise correlations among the various indices.

Table 18
Linear correlations between monthly index variables for Sulphur River at Talco
(Correlations greater than 0.7 in boldface)

	P	SPI-6	HI-6	SPI-12	HI-12	N	SPI-24	PDSI	X	RMC	Q
P	1.00	0.39	0.27	0.27	0.23	0.27	0.16	0.43	0.27	0.03	0.51
SPI-6		1.00	0.47	**0.73**	0.67	**0.71**	0.45	**0.84**	**0.75**	0.16	0.39
HI-6			1.00	0.38	0.36	0.38	0.24	0.41	0.40	0.13	0.41
SPI-12				1.00	**0.95**	**0.98**	0.67	**0.80**	**0.89**	0.25	0.26
HI-12					1.00	**0.96**	0.66	**0.76**	**0.90**	0.28	0.25
N						1.00	0.64	**0.78**	**0.89**	0.25	0.26
SPI-24							1.00	0.60	**0.80**	0.38	0.12
PDSI								1.00	**0.86**	0.29	0.39
X									1.00	0.43	0.28
RMC										1.00	0.05
Q											1.00

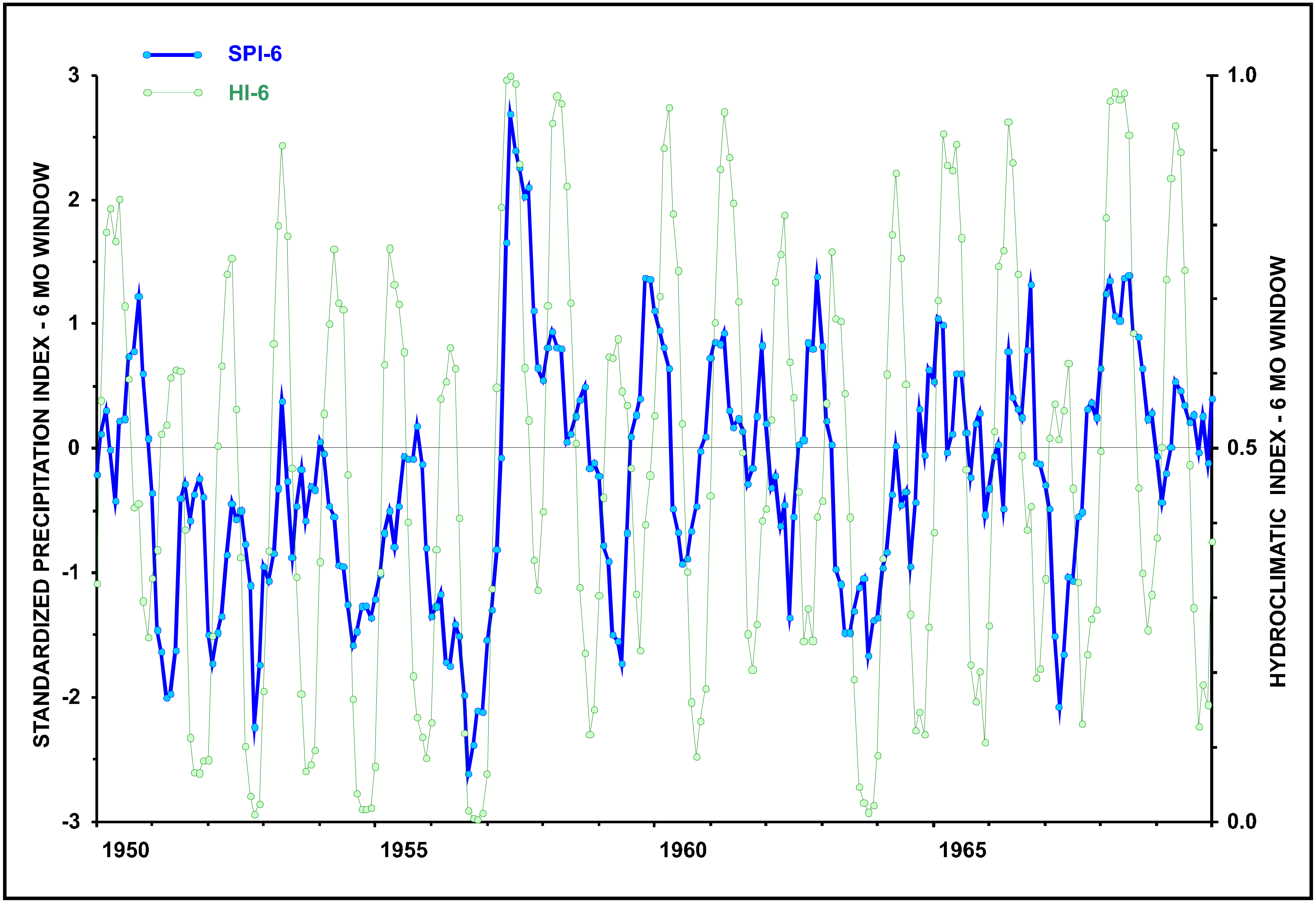

Figure 26 - Sulphur River at Talco short-term memory indices (see Table 17), for 1950-69.

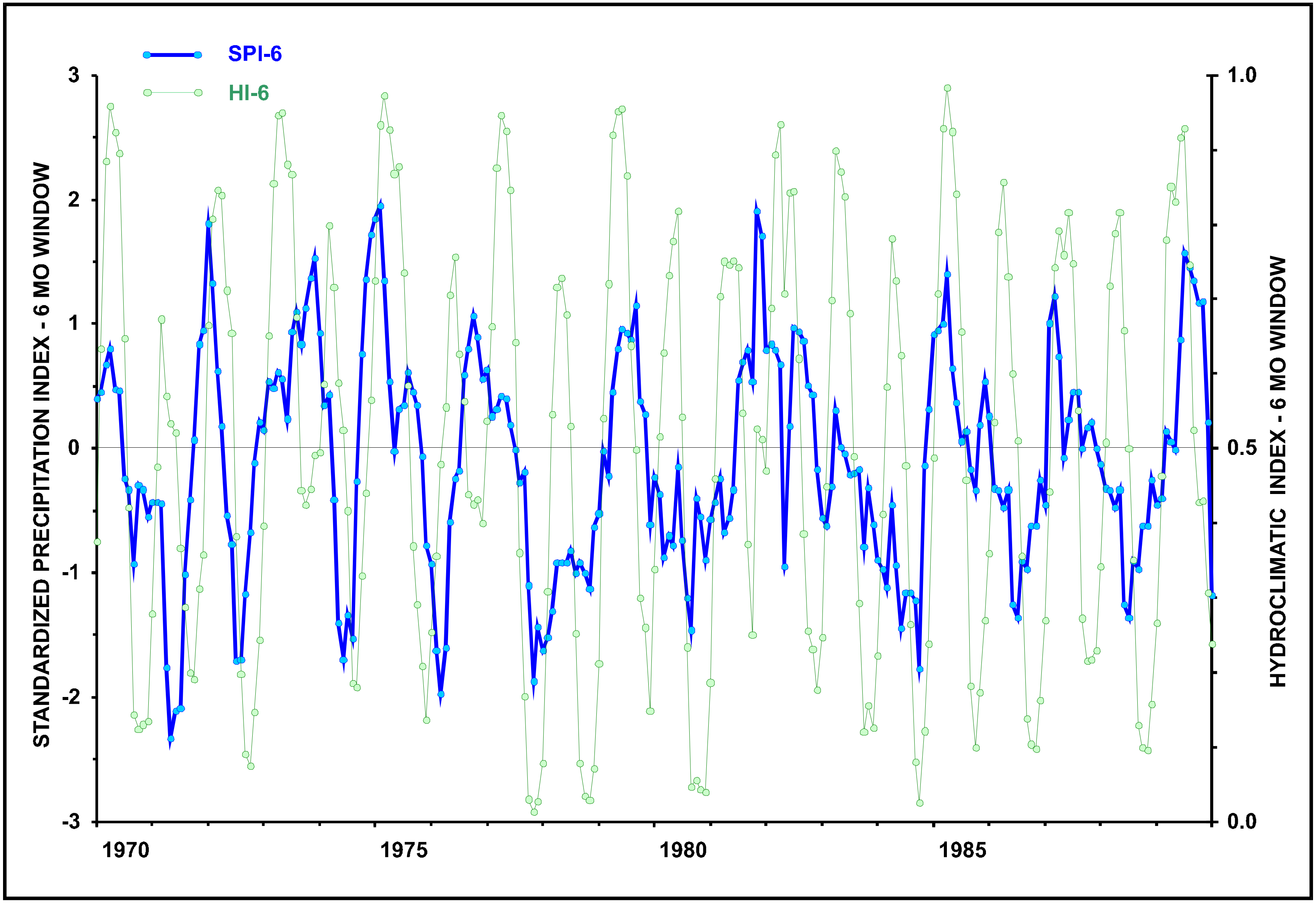

Figure 27 - Sulphur River at Talco short-term memory indices (see Table 17), for 1970-89.

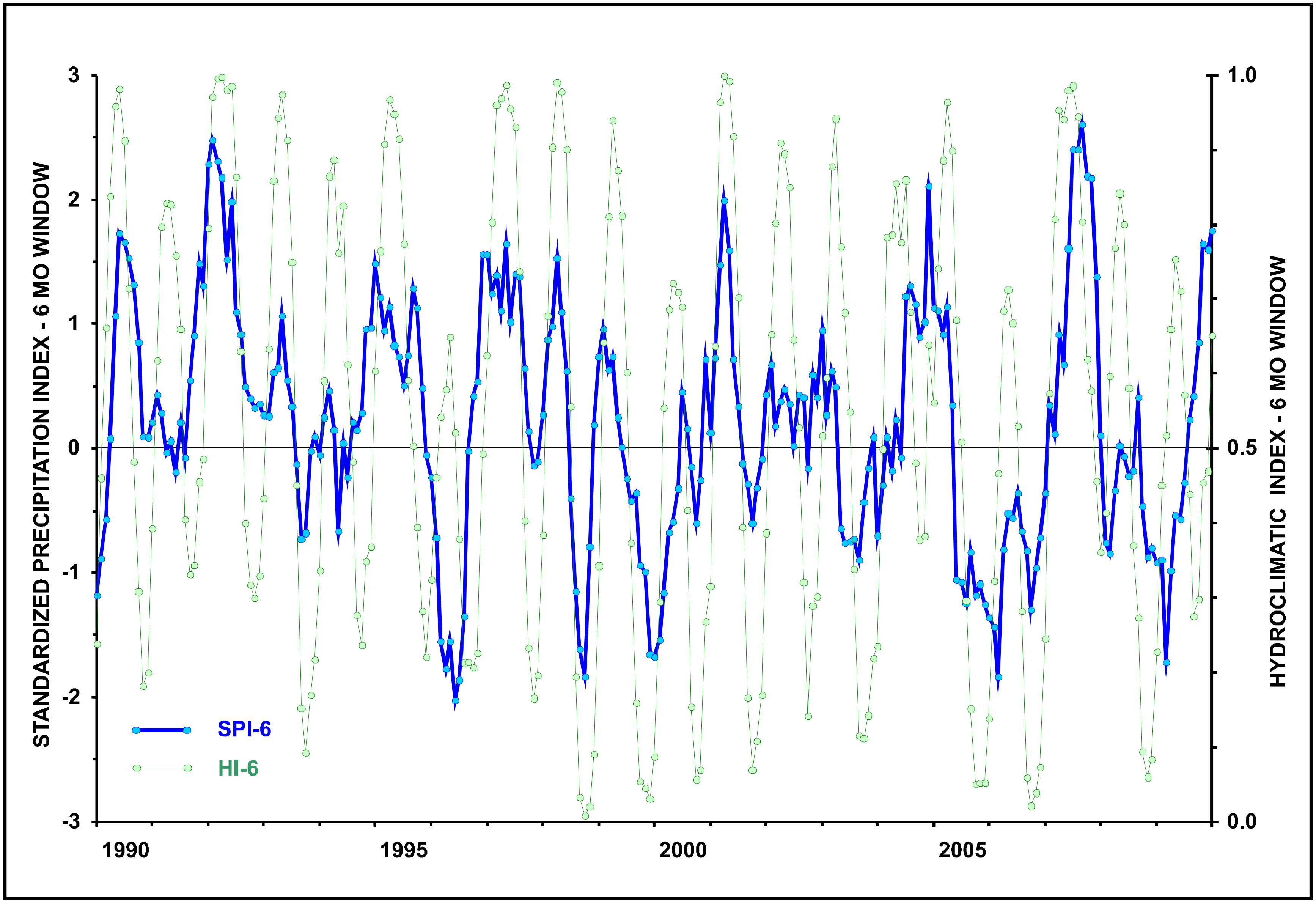

Figure 28 - Sulphur River at Talco short-term memory indices (see Table 17), for 1990-2009.

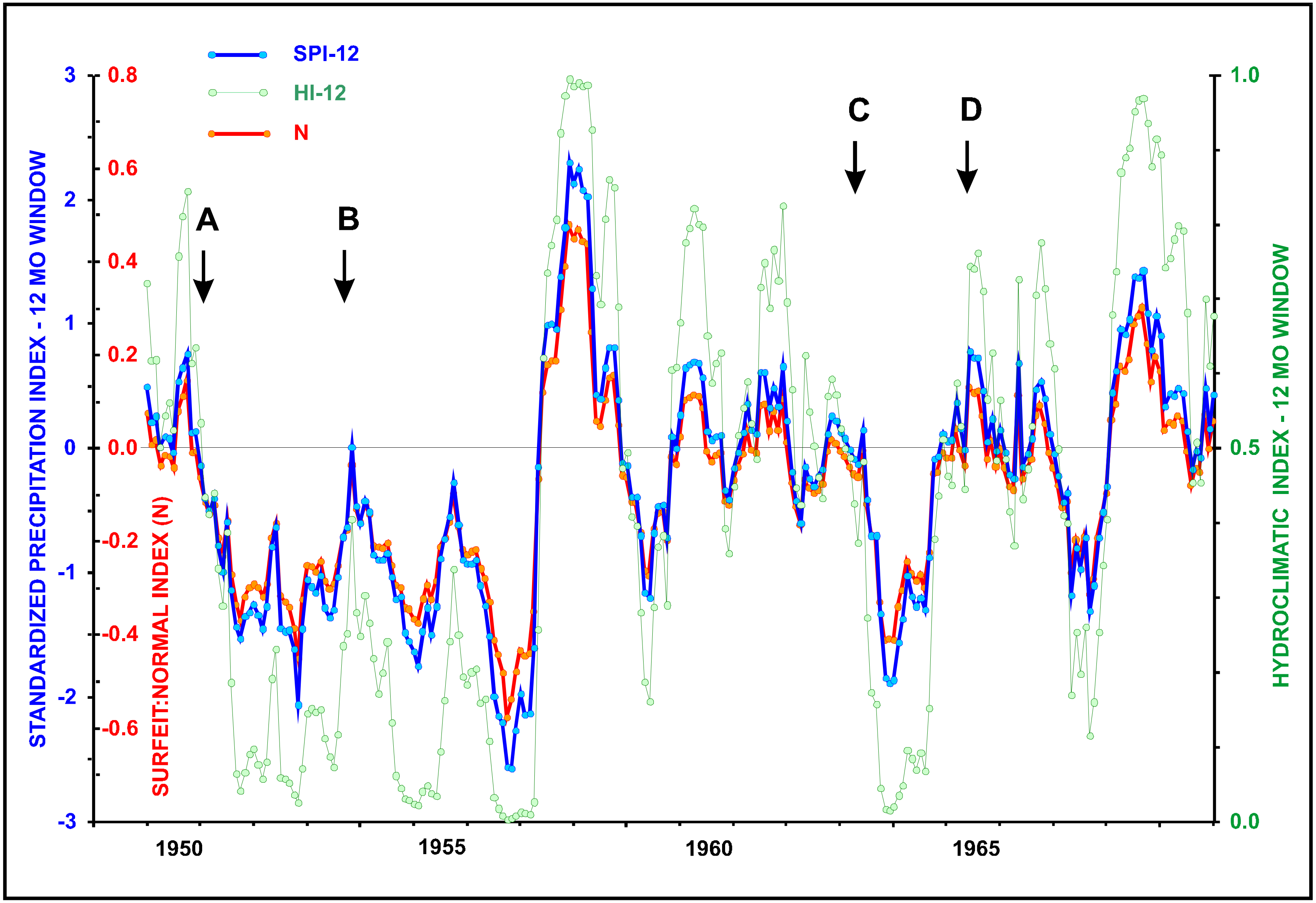

Figure 29 - Sulphur River at Talco moderate (1 year) memory indices (see Table 17), for 1950-1969.

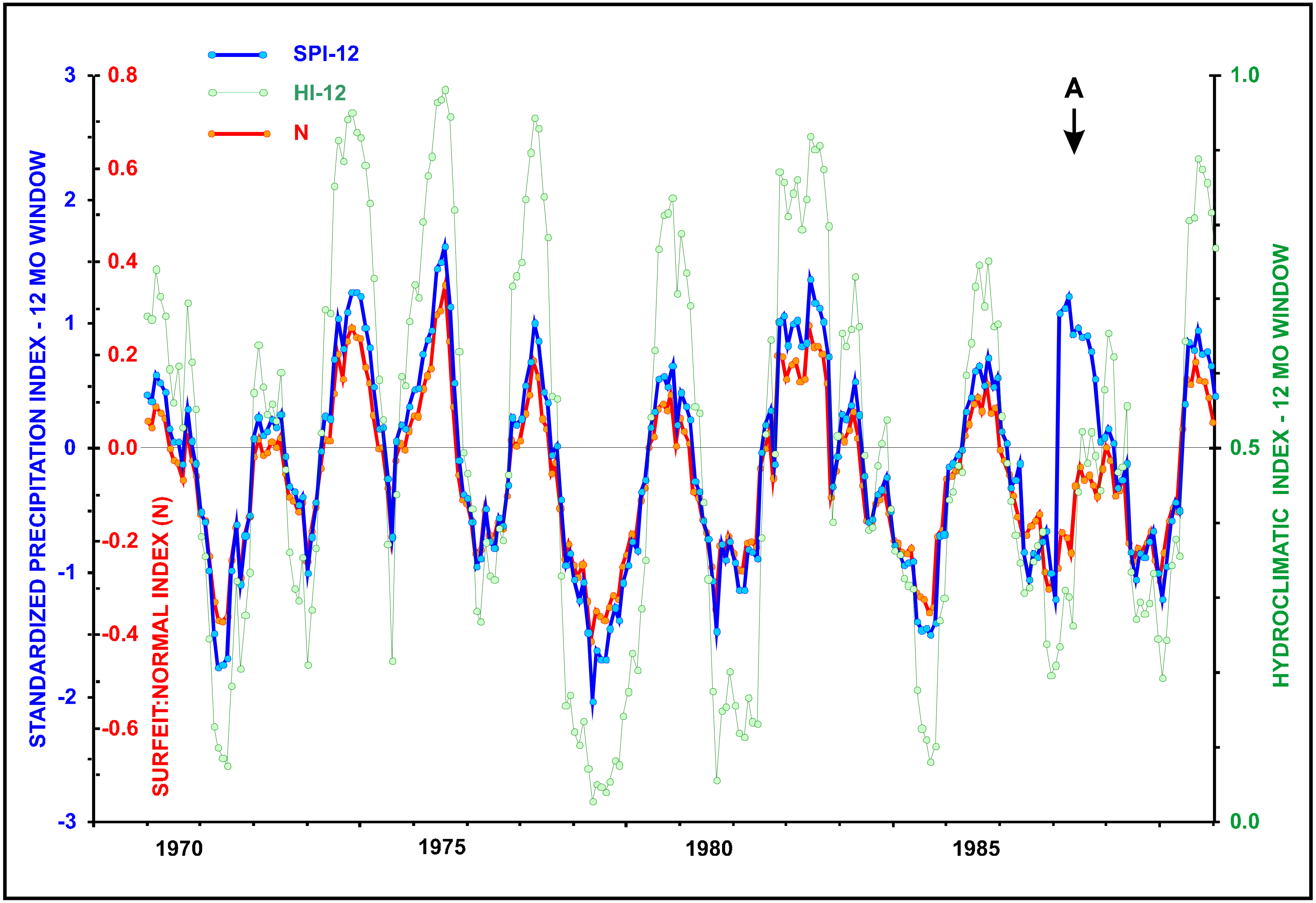

Figure 30 - Sulphur River at Talco moderate (1 year) memory indices (see Table 17), for 1970-1989.

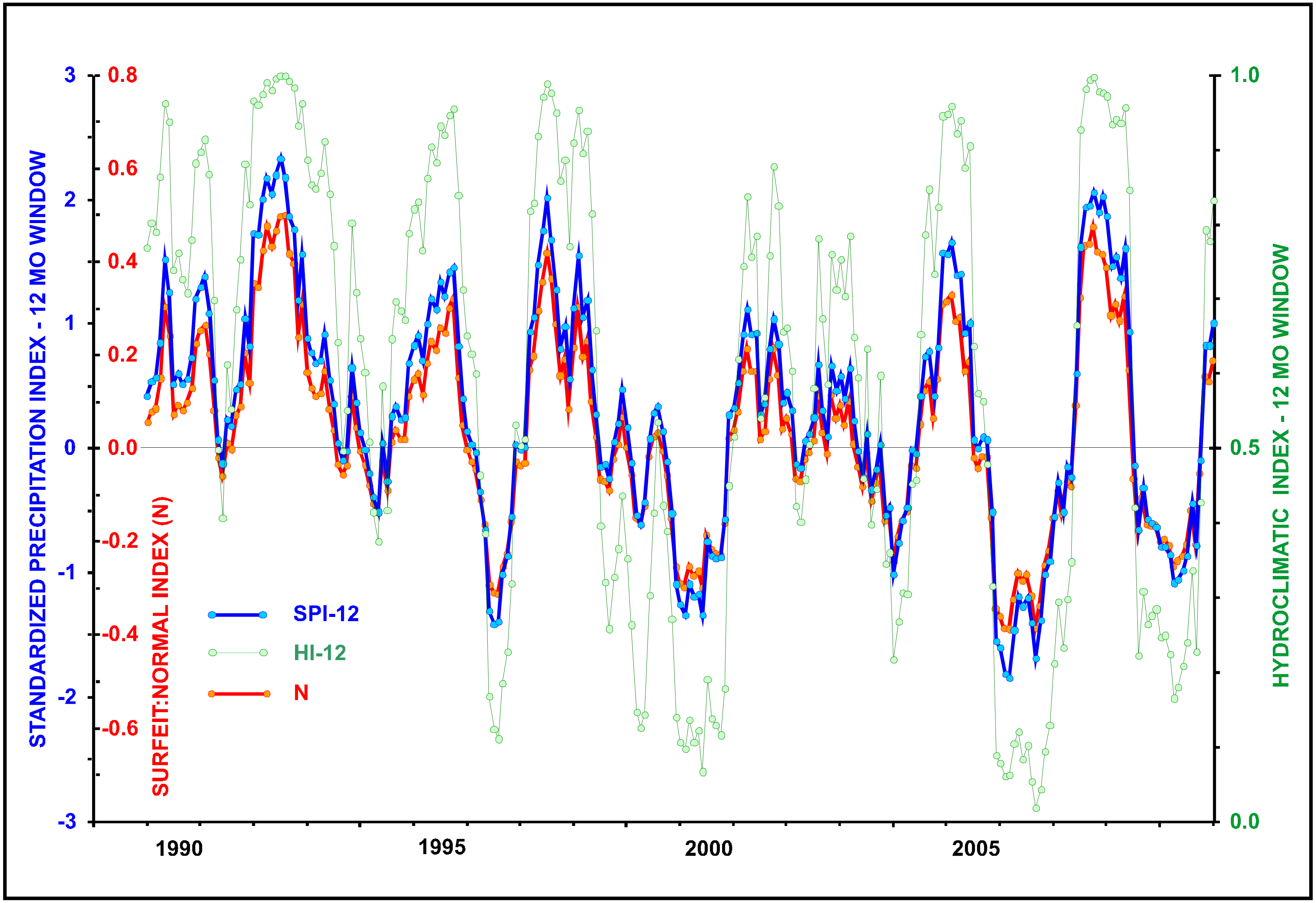

Figure 31 - Sulphur River at Talco moderate (1 year) memory indices (see Table 17), for 1990-2009.

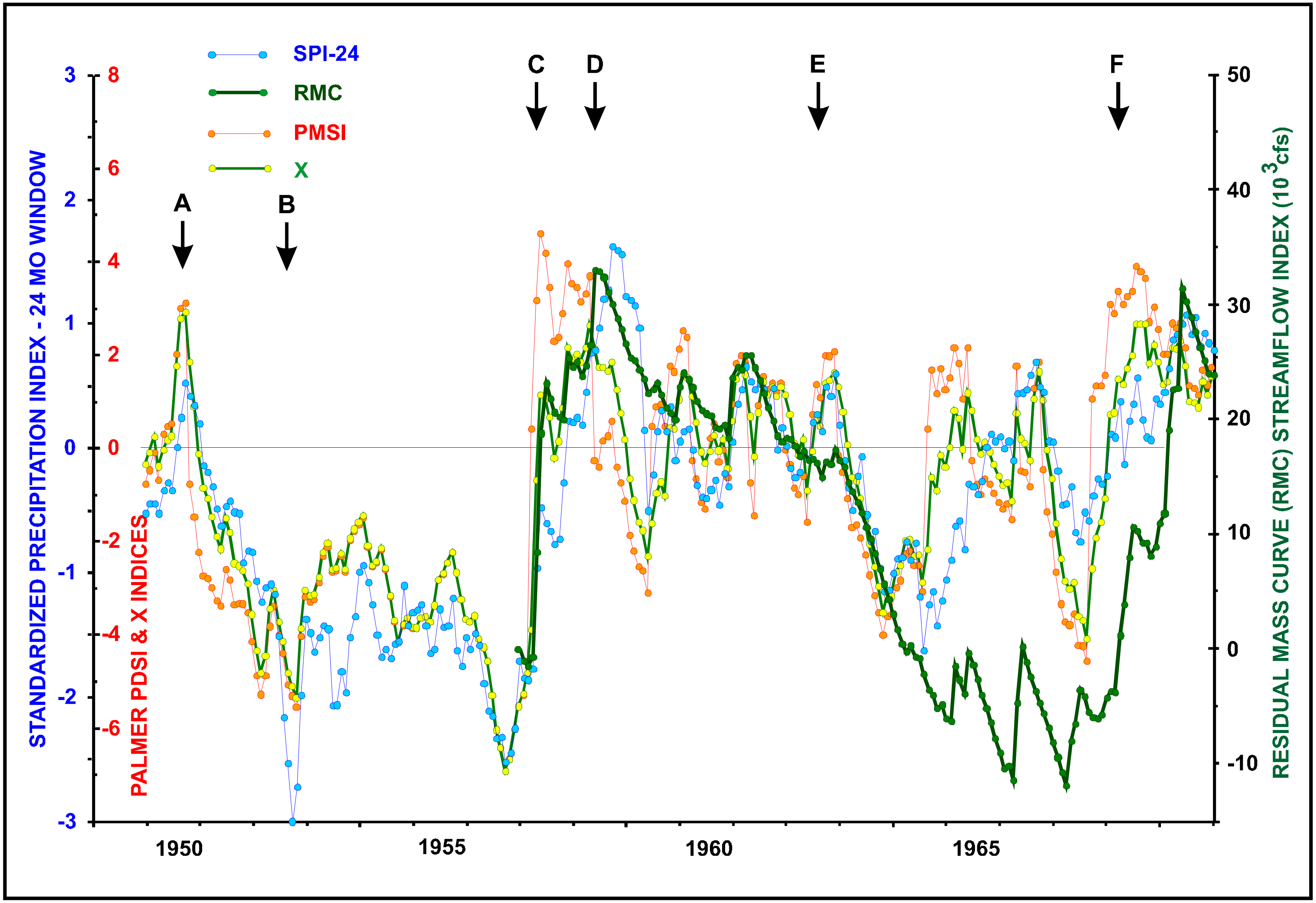

Figure 32 - Sulphur River at Talco long-term memory indices (see Table 17), for 1950-1969.

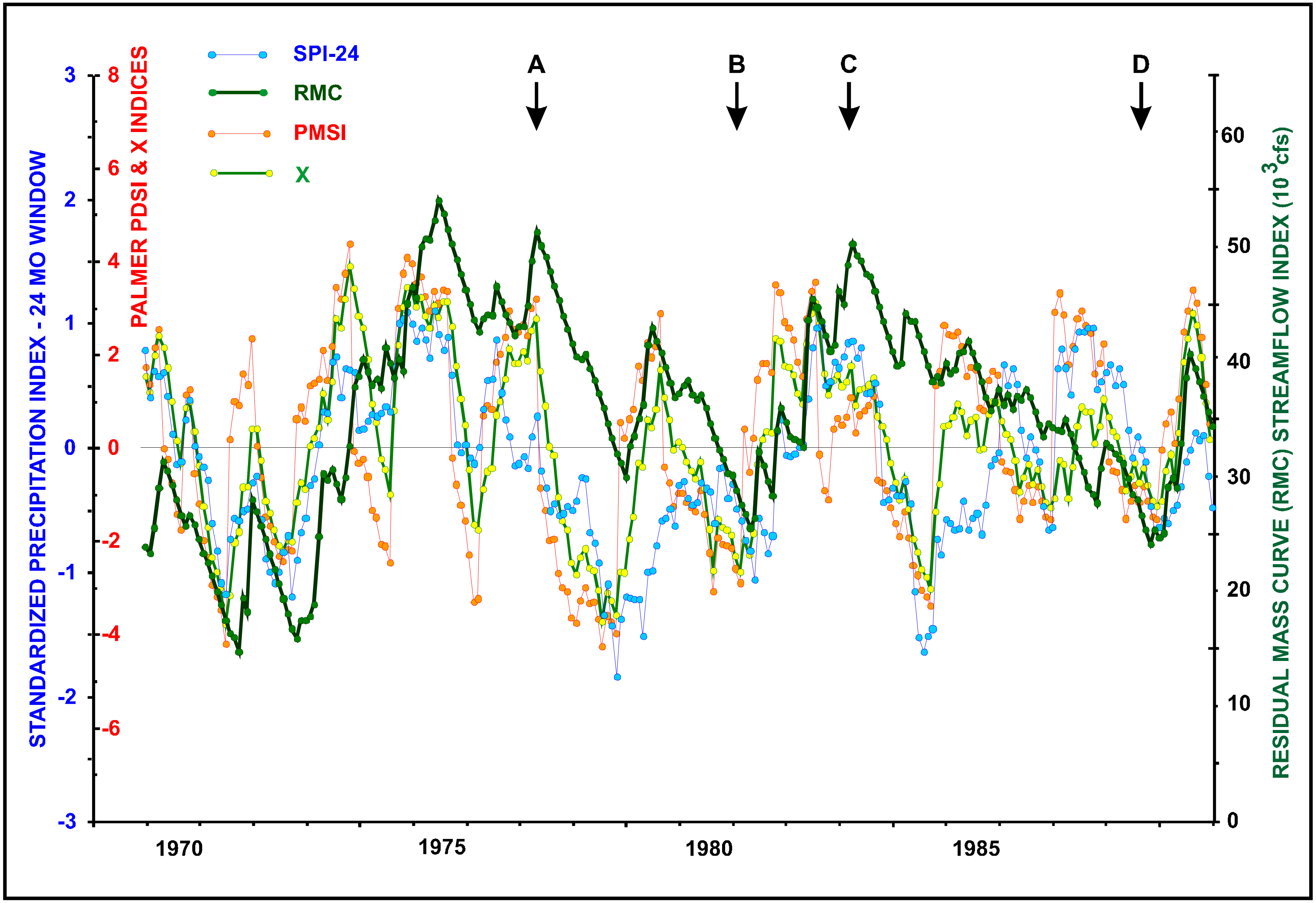

Figure 33 - Sulphur River at Talco long-term memory indices (see Table 17), for 1970-1989.

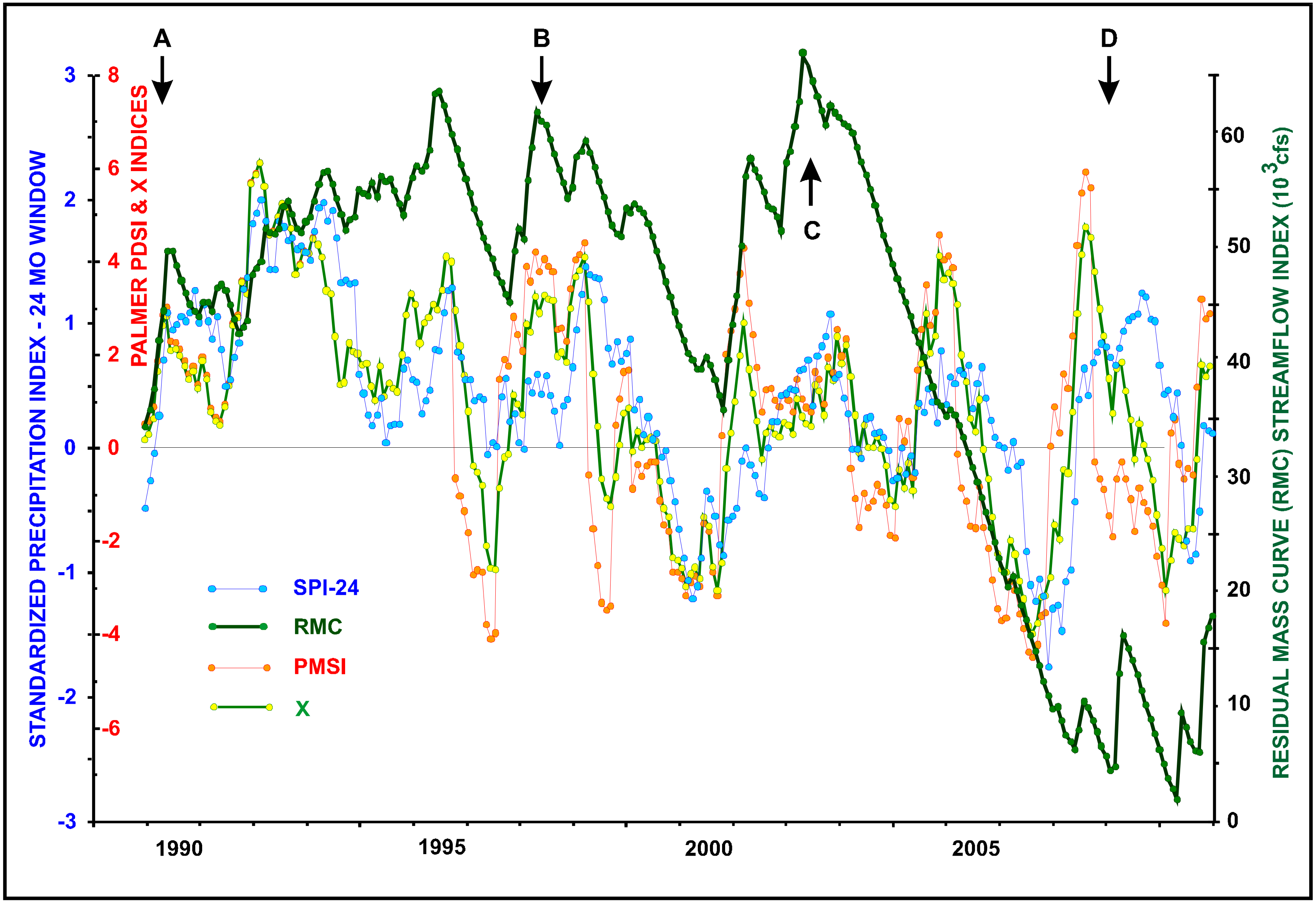

Figure 34 - Sulphur River at Talco long-term memory indices (see Table 17), for 1990-2009.

4.3 Little Cypress Bayou near Jefferson

The Little Cypress flows into Caddo Lake on the Arkansas border. This station is located about 15 miles upstream from the lake backwater, and most of its 1750 km^2 drainage area is contained within the NCDC East Texas division (Division 4). There are no major reservoirs upstream from the gauge site. The record on this gauge starts in June 1946, and is continuous thereafter. Though this river is located in the most water-rich region of the state, during drought it has registered zero monthly flow. The longest sequence of five months of zero flow occurred in the Drought of the 1950's and in 2011. The range on the RMC plot, Figs. 41-43, is 30K with a floating origin. Table 19 summarizes the pairwise correlations among the various indices.

Table 19
Linear correlations between index variables for Little Cypress Bayou near Jefferson
(Correlations greater than 0.7 in boldface)

	P	SPI-6	HI-6	SPI-12	HI-12	N	SPI-24	PDSI	X	RMC	Q
P	1.00	0.12	0.27	0.10	0.30	0.32	0.04	0.51	0.35	-0.04	0.48
SPI-6		1.00	0.48	**0.76**	**0.72**	**0.75**	0.50	**0.84**	**0.80**	0.02	0.45
HI-6			1.00	0.41	0.40	0.41	0.27	0.43	0.42	0.06	0.62
SPI-12				1.00	**0.97**	**0.99**	**0.72**	**0.82**	**0.91**	0.11	0.40
HI-12					1.00	**0.98**	**0.71**	**0.80**	**0.92**	0.13	0.38
N						1.00	**0.71**	**0.81**	**0.91**	0.15	0.39
SPI-24							1.00	0.61	**0.82**	0.23	0.27
PDSI								1.00	**0.87**	0.01	0.47
X									1.00	0.16	0.41
RMC										1.00	0.06
Q											1.00

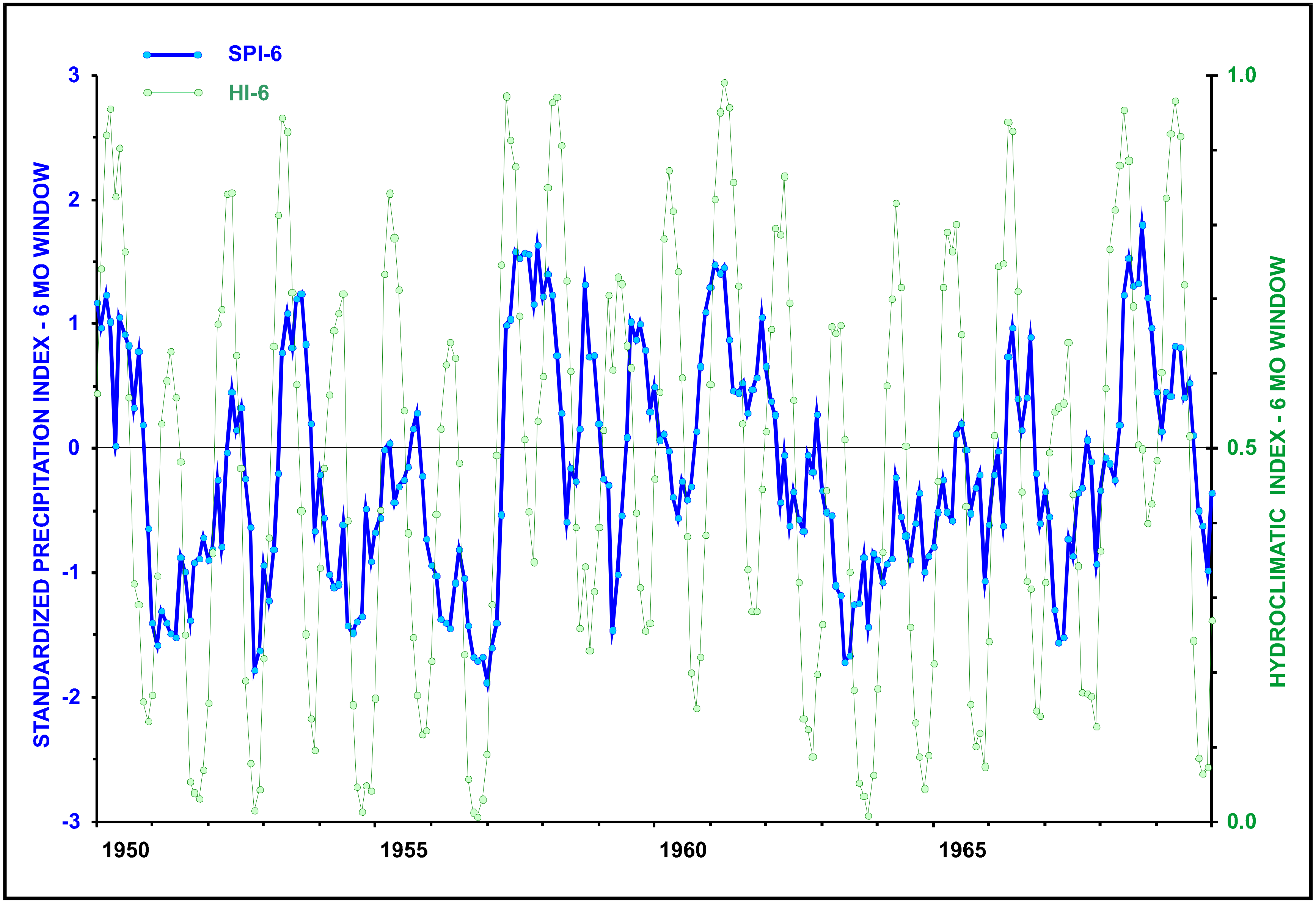

Figure 35 - Little Cypress Bayou near Jefferson short-term memory indices (see Table 17), for 1950-69.

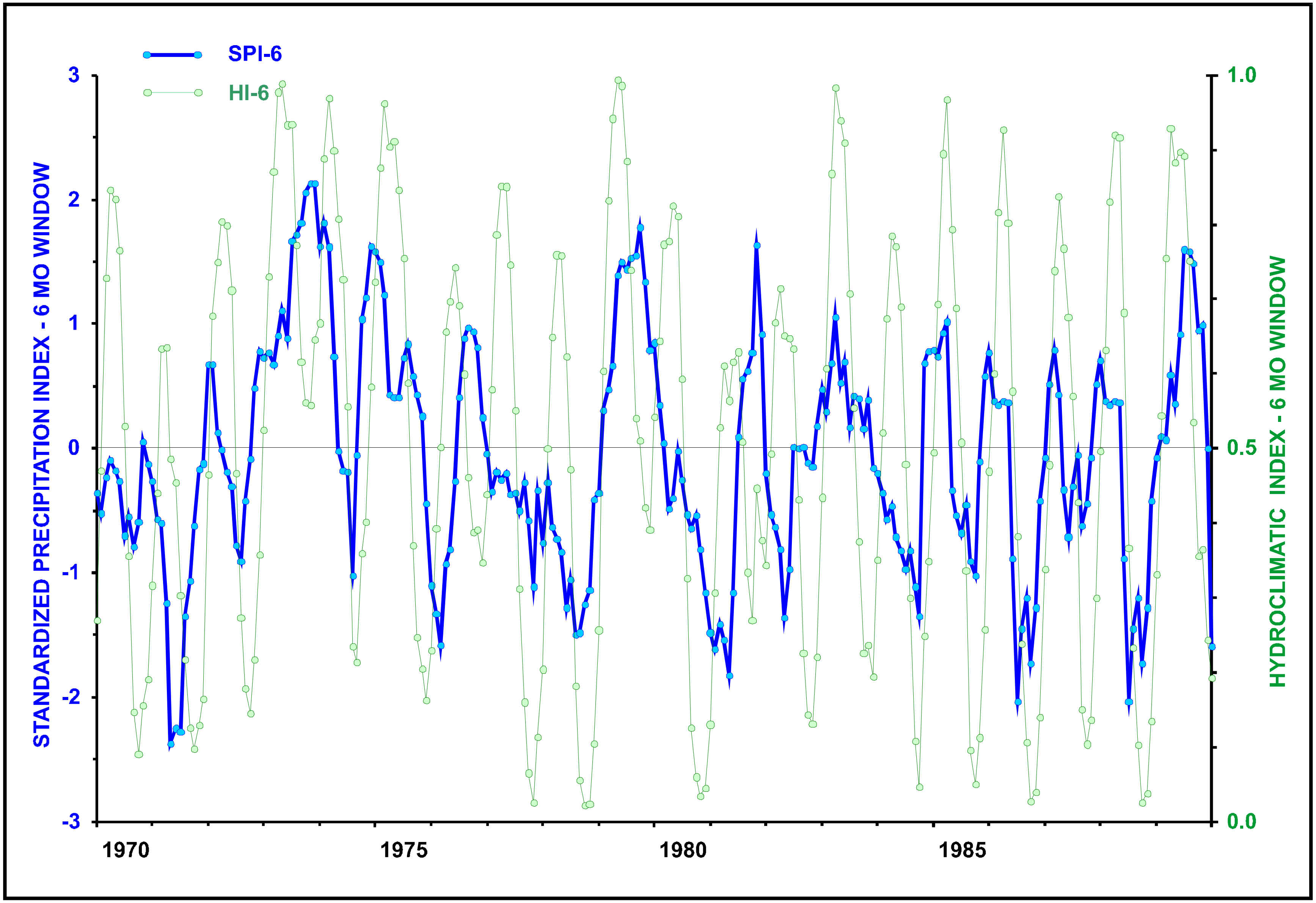

Figure 36 - Little Cypress Bayou near Jefferson short-term memory indices (see Table 17), for 1970-89.

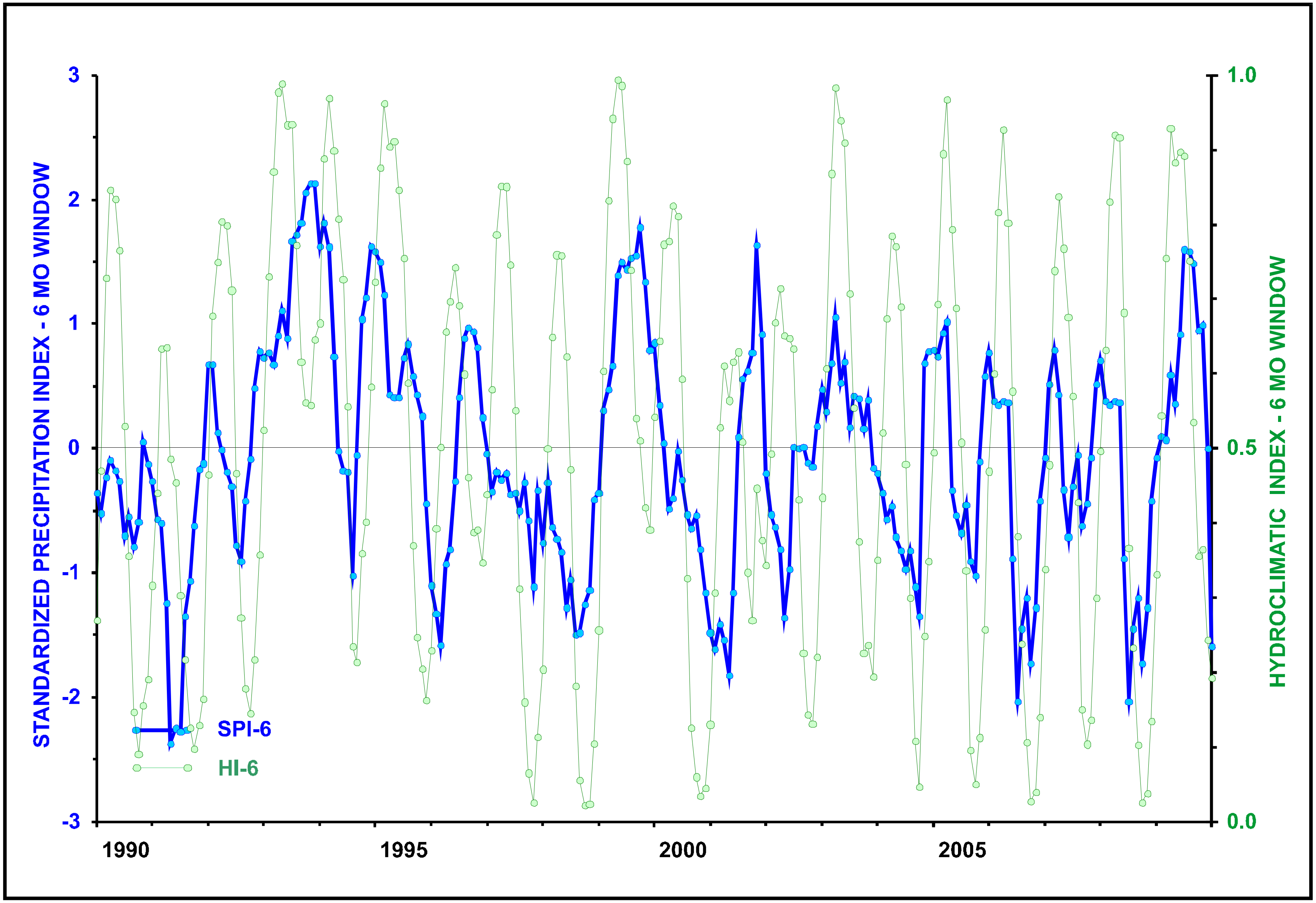

Figure 37 - Little Cypress Bayou near Jefferson short-term memory indices (see Table 17), for 1990-2009.

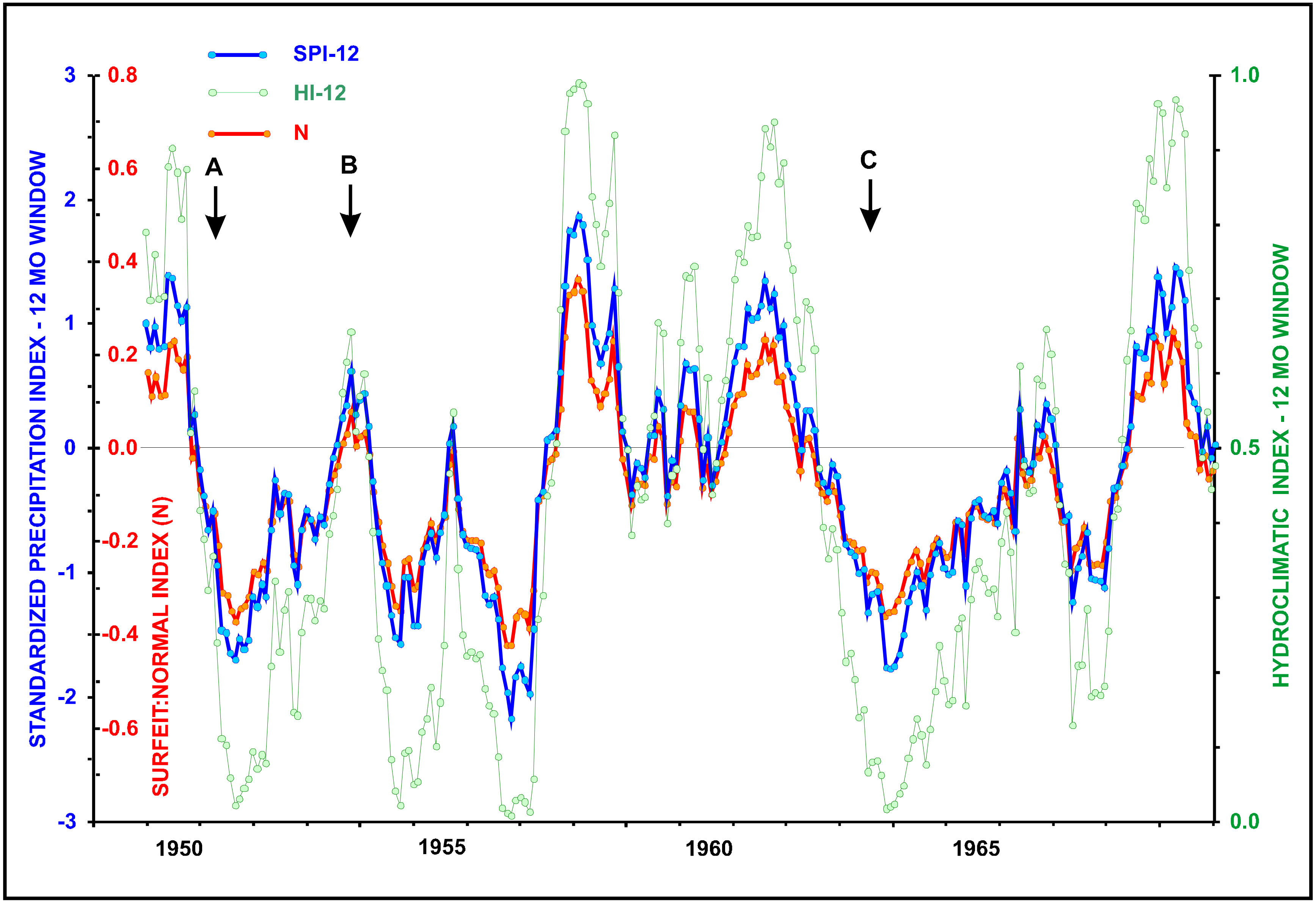

Figure 38 - Little Cypress Bayou near Jefferson moderate (1 year) memory indices (see Table 17), for 1950-1969.

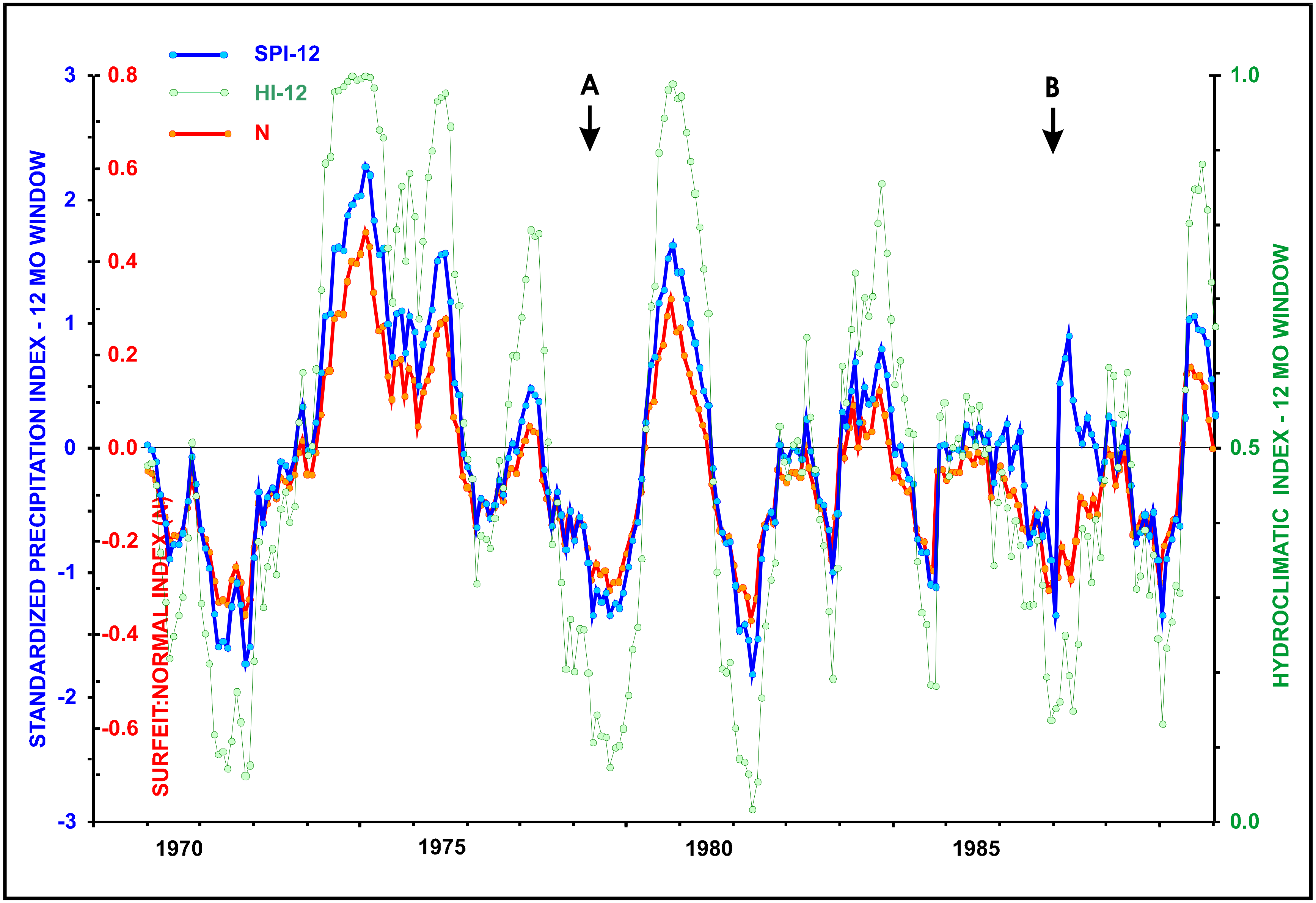

Figure 39 - Little Cypress Bayou near Jefferson moderate (1 year) memory indices (see Table 17), for 1970-1989.

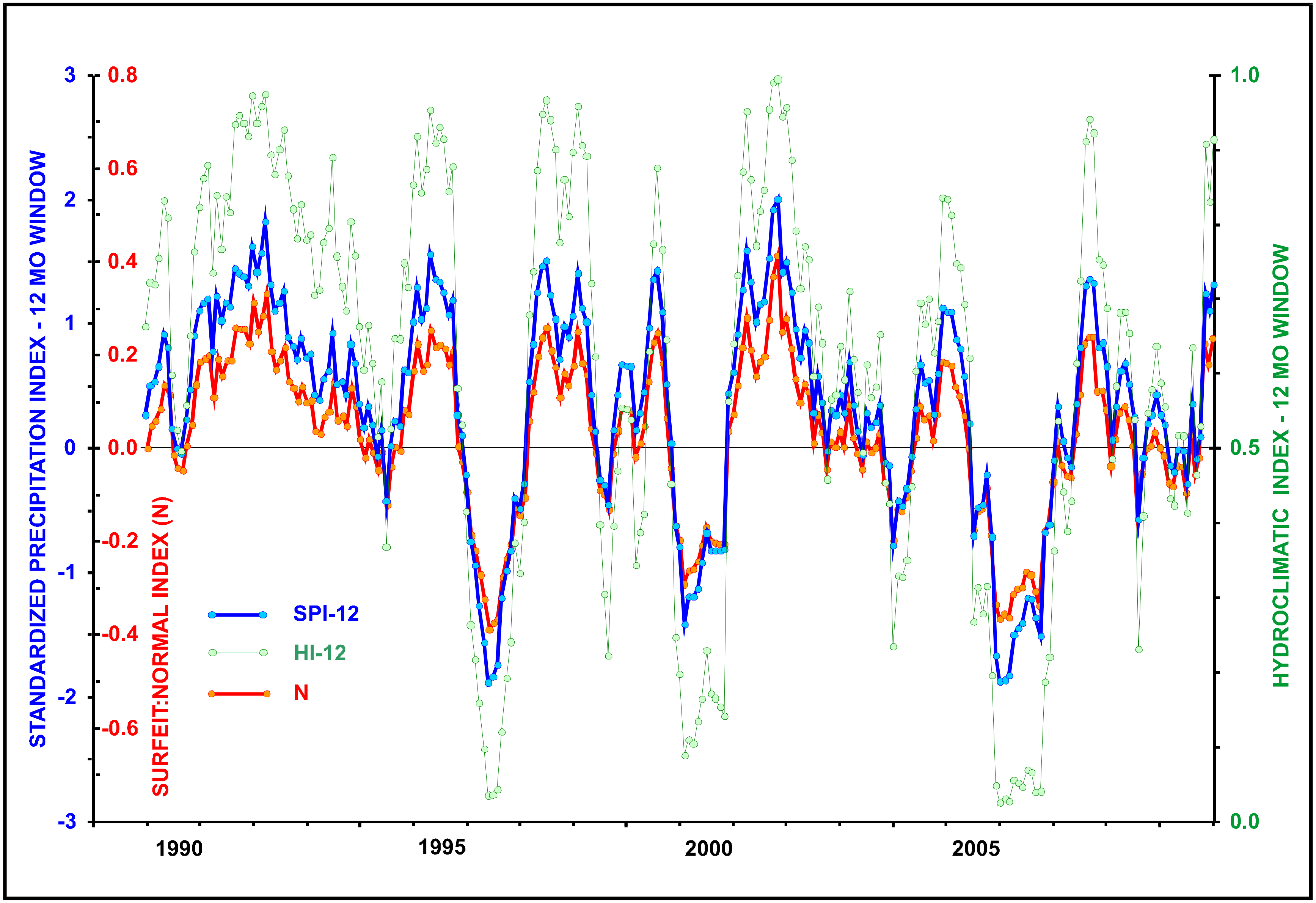

Figure 40 - Little Cypress Bayou near Jefferson moderate (1 year) memory indices (see Table 17), for 1990-2009.

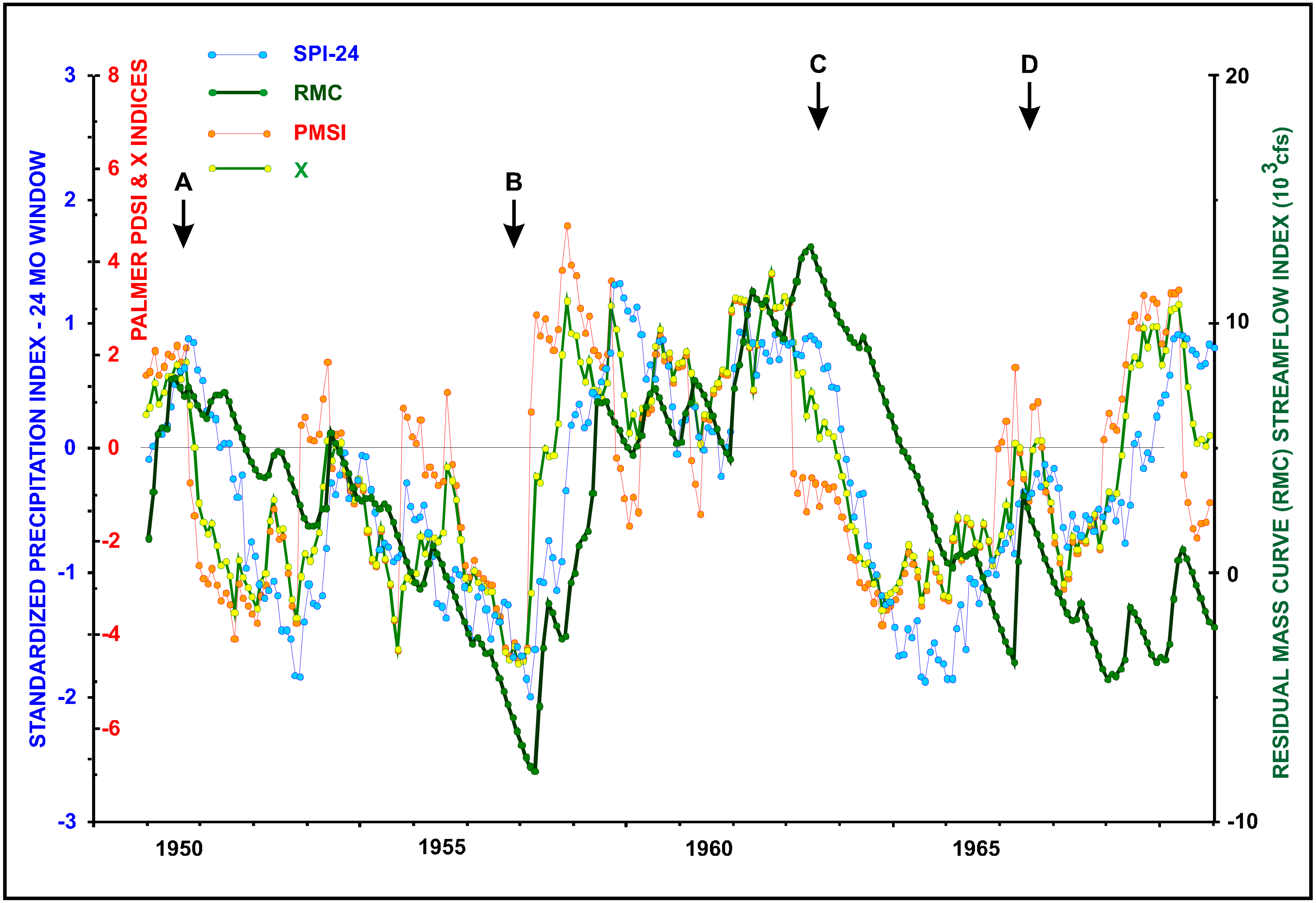

Figure 41 - Little Cypress Bayou near Jefferson long-term memory indices (see Table 17), for 1950-1969.

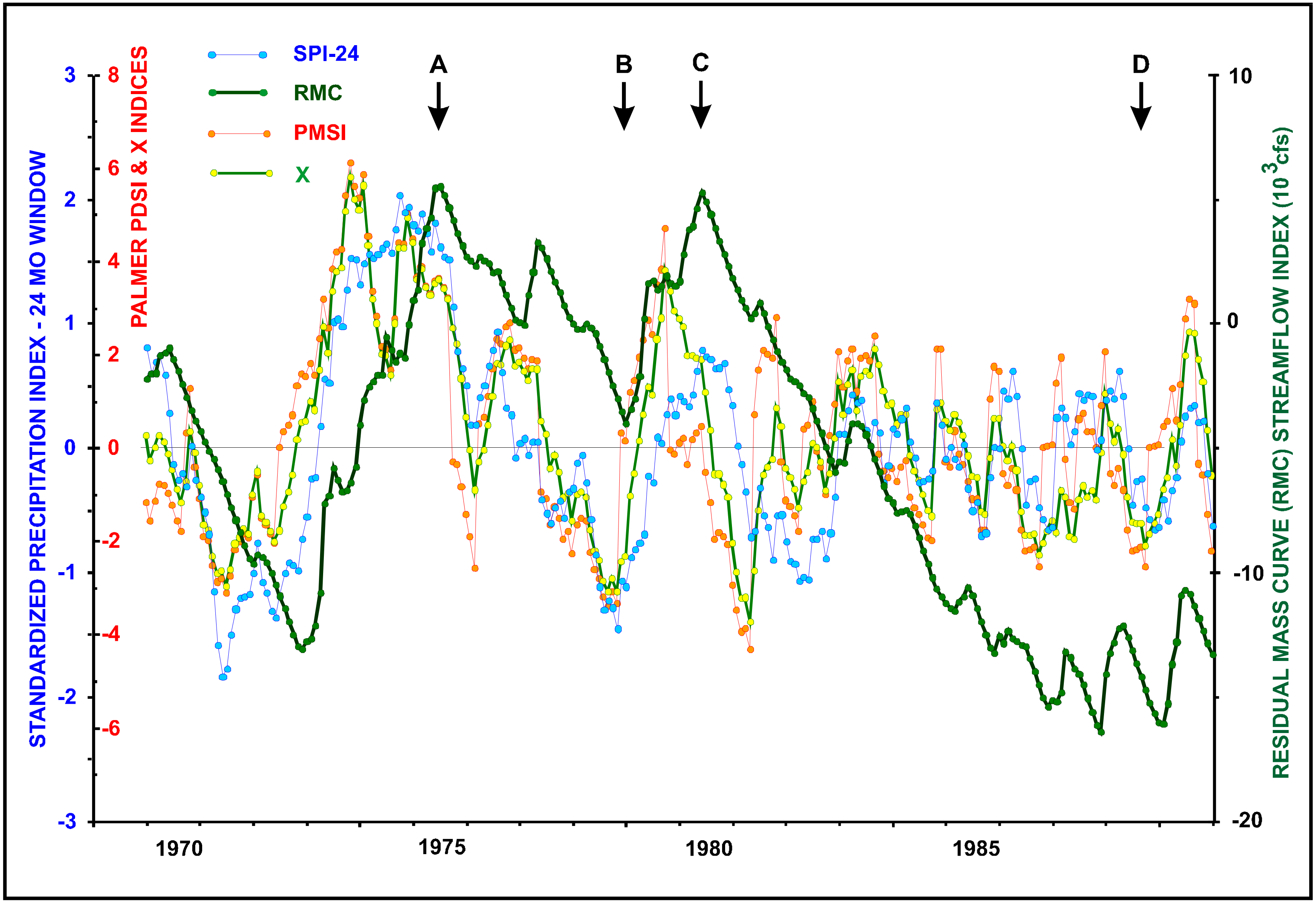

Figure 42 - Little Cypress Bayou near Jefferson long-term memory indices (see Table 17), for 1970-1989.

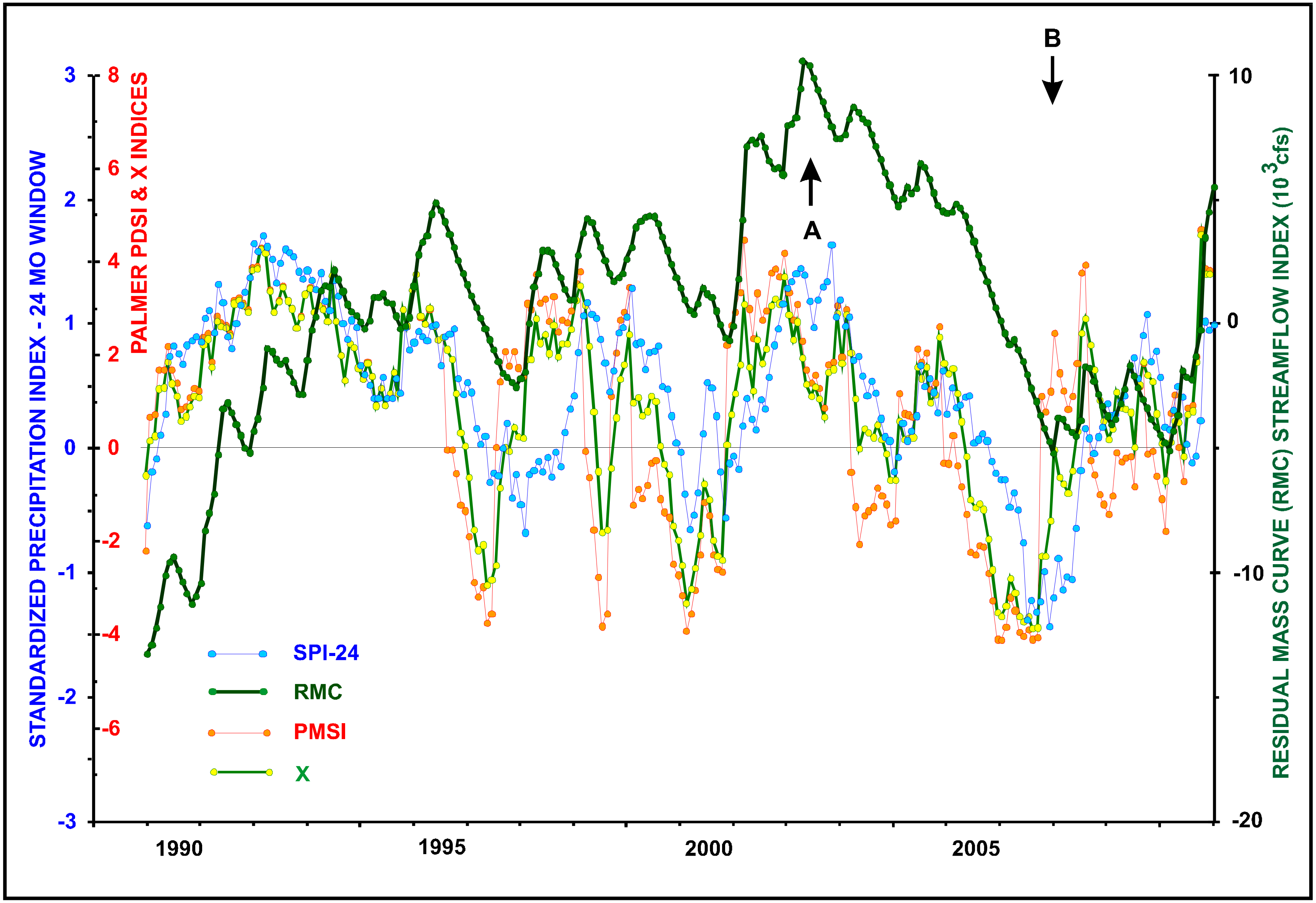

Figure 43 - Little Cypress Bayou near Jefferson long-term memory indices (see Table 17), for 1990-2009.

4.4 Trinity River at Oakwood

Though the Oakwood gauge is located within the East Texas climatic division (Division 4), almost the entirety of the watershed is in the North Central division (Division 3), so that division is used as the source for meteorological and moisture condition data. There are numerous reservoirs in the upper Trinity basin, but this gauge is a considerable distance downstream. The watershed at the gauge is 33,250 km^2, and is one of the largest watersheds represented in this selection of test sites. This gauge has a continuous record since October 1923. USGS has never registered a zero monthly flow here; the lowest daily reading of 7.8 cfs occurred in July 1924.

Because the climatic division for this gauge is the same as the Sulphur (Section 4.1, above), the climatological indices are the same as shown in Figs. 26-34, so there is no need to repeat them here. However, the RMC index will obviously be different at this gauge, so only the indices with long-term memory are plotted in Figures 44-46. The range on the RMC plot is 500K (note change of axis unit) with a floating origin. Table 20 summarizes the pairwise correlations among the various indices.

Table 20
Linear correlations between monthly index variables for Trinity River at Oakwood
(Correlations greater than 0.7 in boldface)

	P	SPI-6	HI-6	SPI-12	HI-12	N	SPI-24	PDSI	X	RMC	Q
P	1.00	0.39	0.27	0.27	0.23	0.27	0.16	0.43	0.27	-0.03	0.45
SPI-6		1.00	0.48	**0.73**	0.67	**0.71**	0.45	**0.84**	**0.75**	-0.01	0.55
HI-6			1.00	0.38	0.37	0.38	0.24	0.41	0.40	-0.02	0.55
SPI-12				1.00	**0.95**	**0.98**	0.67	**0.80**	**0.89**	0.02	0.43
HI-12					1.00	**0.96**	0.66	**0.76**	**0.90**	-0.04	0.41
N						1.00	0.64	**0.78**	**0.89**	0.02	0.43
SPI-24							1.00	0.60	**0.80**	0.04	0.30
PDSI								1.00	**0.86**	-0.09	0.51
X									1.00	-0.01	0.46
RMC										1.00	0.02
Q											1.00

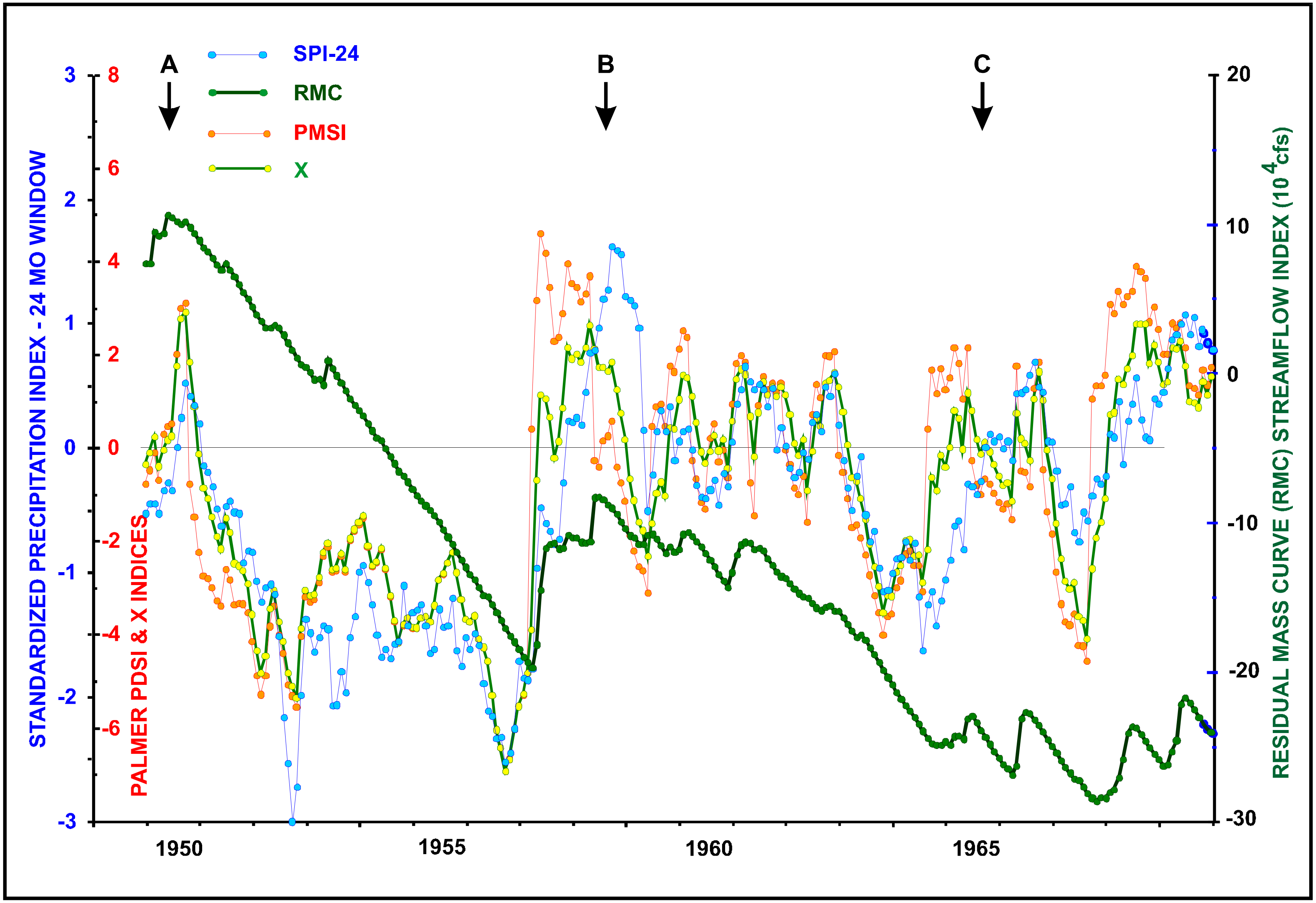

Figure 44 - Trinity River at Oakwood long-term memory indices (see Table 17), for 1950-1969

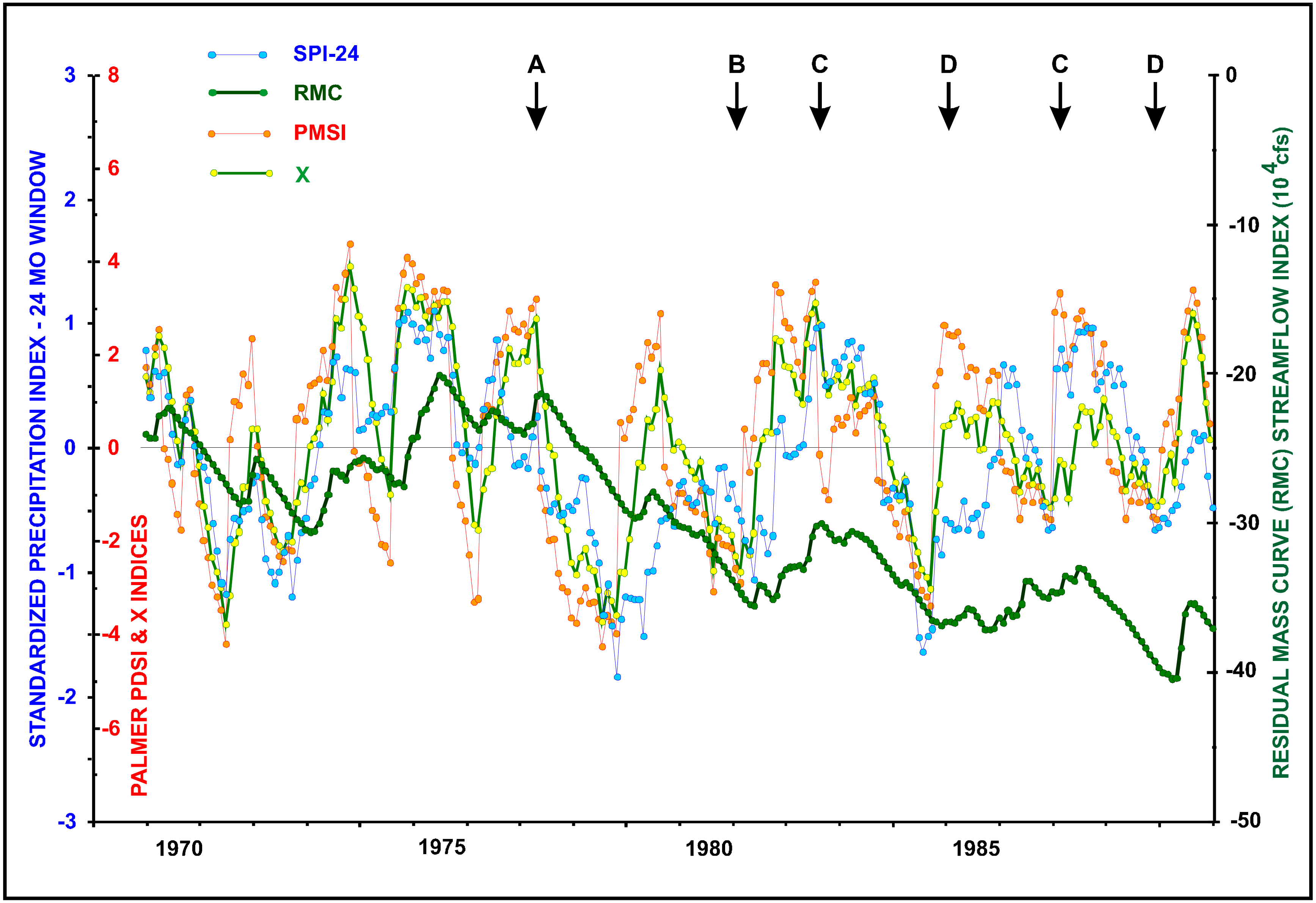

Figure 45 - Trinity River at Oakwood long-term memory indices (see Table 17), for 1970-1989

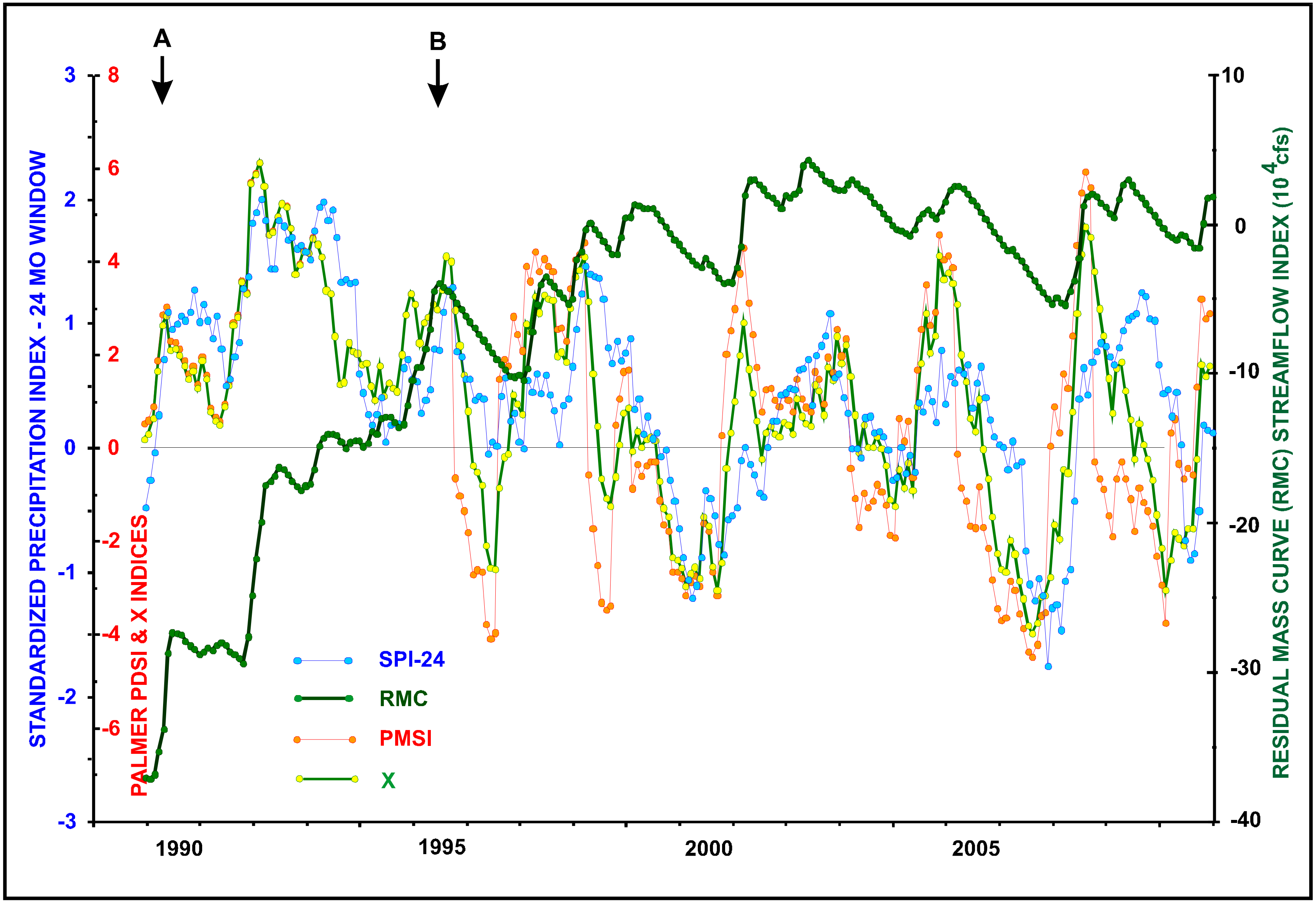

Figure 46 - Trinity River at Oakwood long-term memory indices (see Table 17), for 1990-2009

4.5 Brazos River at Seymour

The Seymour gauge is located on the eastern margin of the Low Rolling Plains climatic division (Division 2). The headwaters of the river lie within the High Plains division to the west, and the river traverses the width of the Low Rolling Plains division. Relatively little runoff is contributed from the High Plains, however, so for this analysis, Division 2 climatology is assumed to prevail. Though this gauge is high in the basin, it drains 15,500 km^2 (and much more noncontributing). There are no major reservoirs upstream. The gauge record is continuous since December 1923. Given its western location, it is no surprise that USGS occasionally records a zero monthly flow, the longest sequence occurring in 2011. The range of the RMC plot, Figs. 53-55, is 30K, with a floating origin. Table 21 summarizes the pairwise correlations among the various indices.

Table 21
Linear correlations between monthly index variables for Brazos River at Seymour
(Correlations greater than 0.7 in boldface)

	P	SPI-6	HI-6	SPI-12	HI-12	N	SPI-24	PDSI	X	RMC	Q
P	1.00	0.33	0.15	0.23	0.19	0.23	0.16	0.38	0.22	0.08	0.65
SPI-6		1.00	0.46	**0.70**	0.62	0.68	0.45	**0.78**	0.67	0.23	0.32
HI-6			1.00	0.36	0.35	0.36	0.25	0.39	0.35	0.10	0.13
SPI-12				1.00	**0.91**	**0.97**	0.65	**0.78**	**0.83**	0.32	0.20
HI-12					1.00	**0.93**	0.64	**0.74**	**0.85**	0.36	0.16
N						1.00	0.63	**0.77**	**0.84**	0.33	0.20
SPI-24							1.00	0.61	**0.78**	0.42	0.10
PDSI								1.00	**0.86**	0.36	0.26
X									1.00	0.41	0.15
RMC										1.00	0.05
Q											1.00

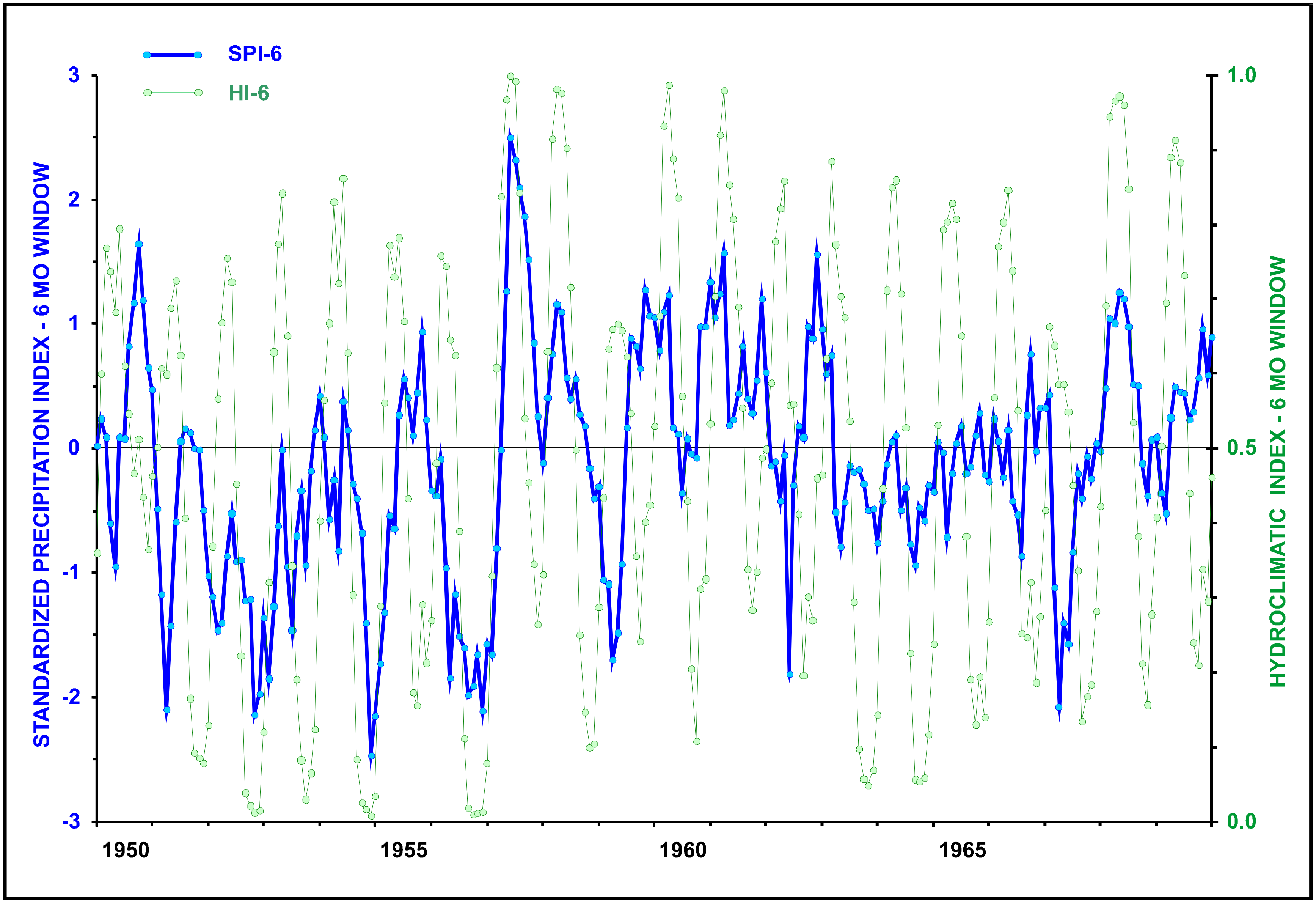

Figure 47 - Brazos River at Seymour short-term memory indices (see Table 17), for 1950-69

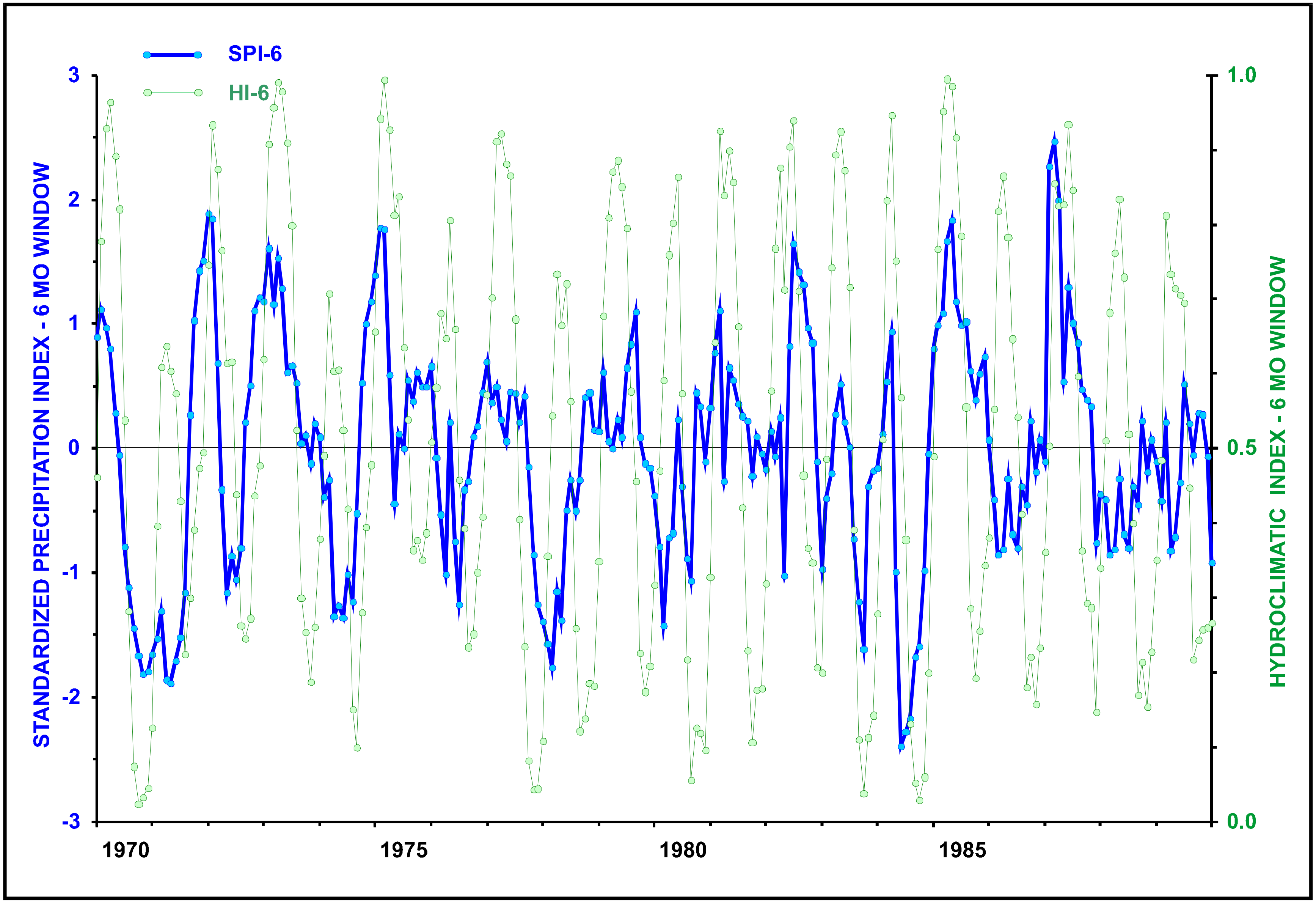

Figure 48 - Brazos River at Seymour short-term memory indices (see Table 17), for 1970-89

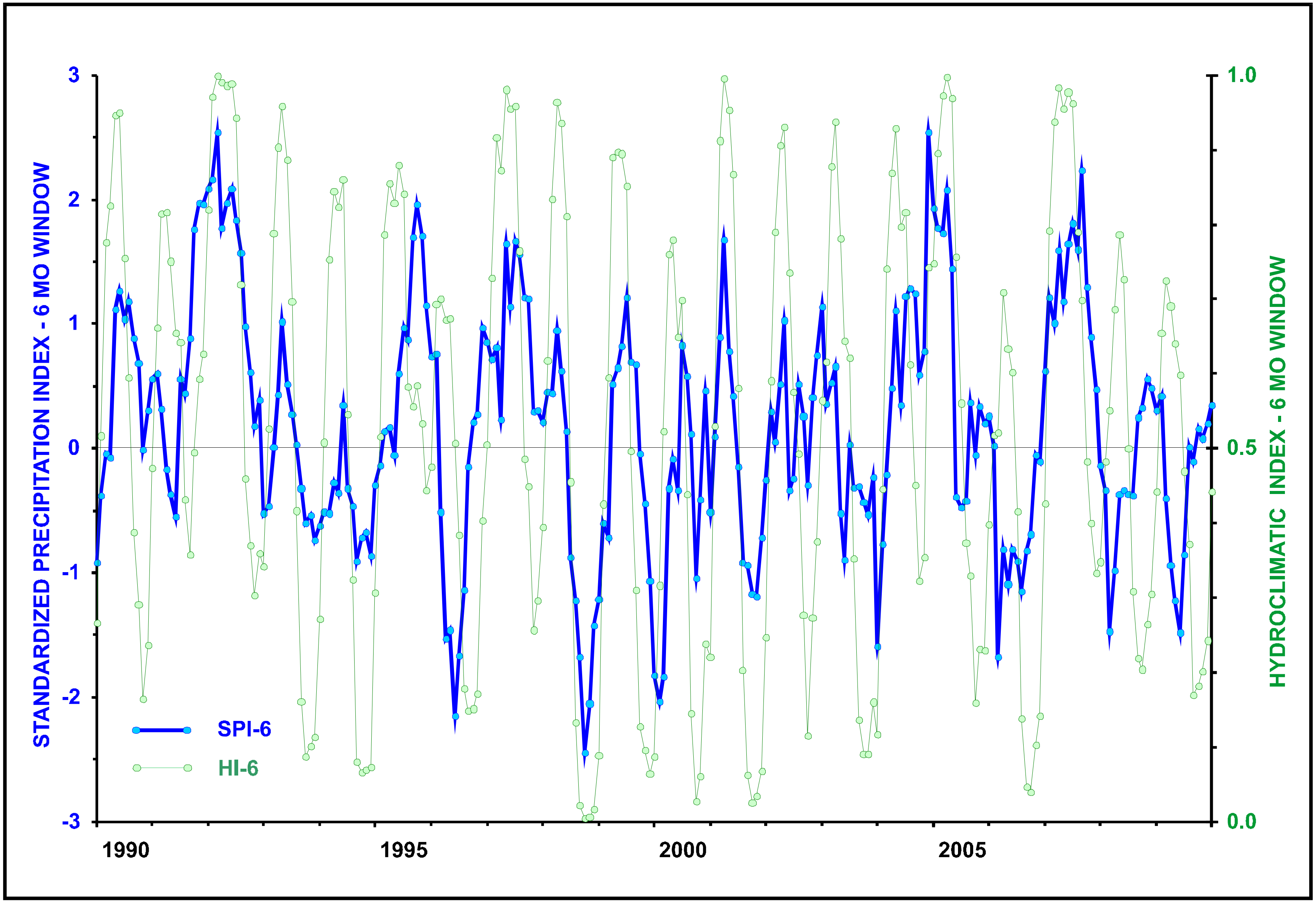

Figure 49 - Brazos River at Seymour short-term memory indices (see Table 17), for 1990-2009

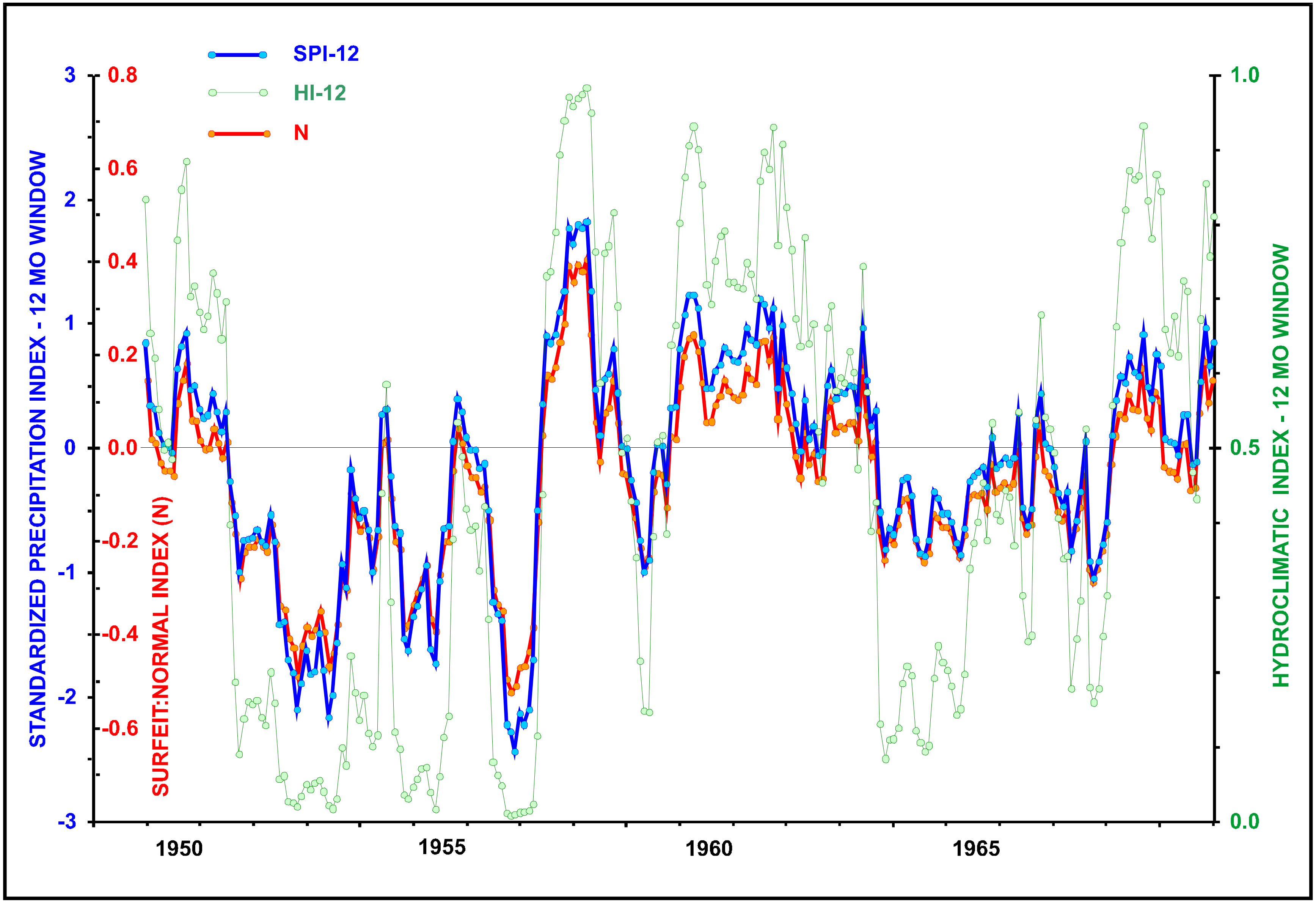

Figure 50 - Brazos River near Seymour moderate (1 year) memory indices (see Table 17), for 1950-1969

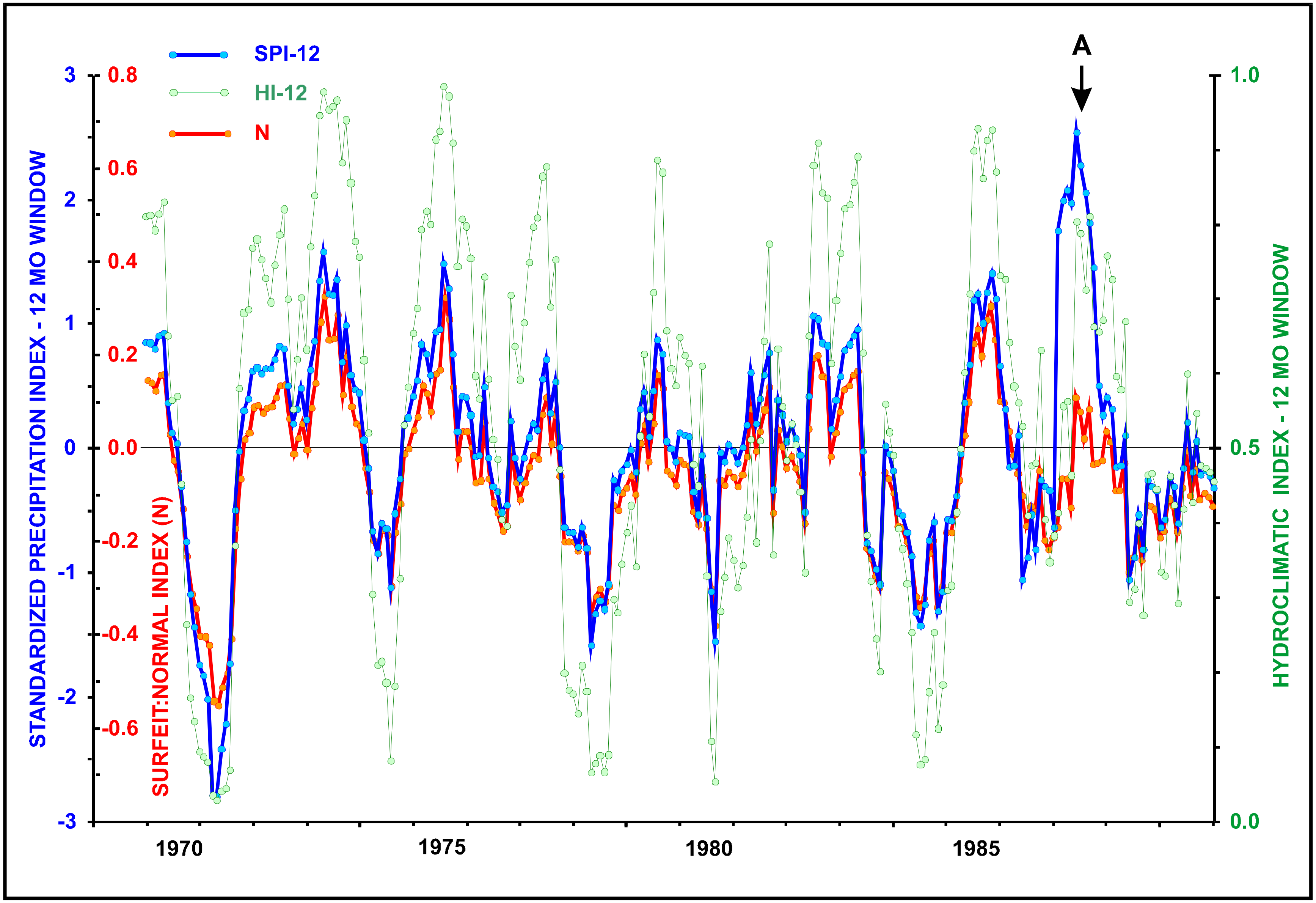

Figure 51 - Brazos River near Seymour moderate (1 year) memory indices (see Table 17), for 1970-1989

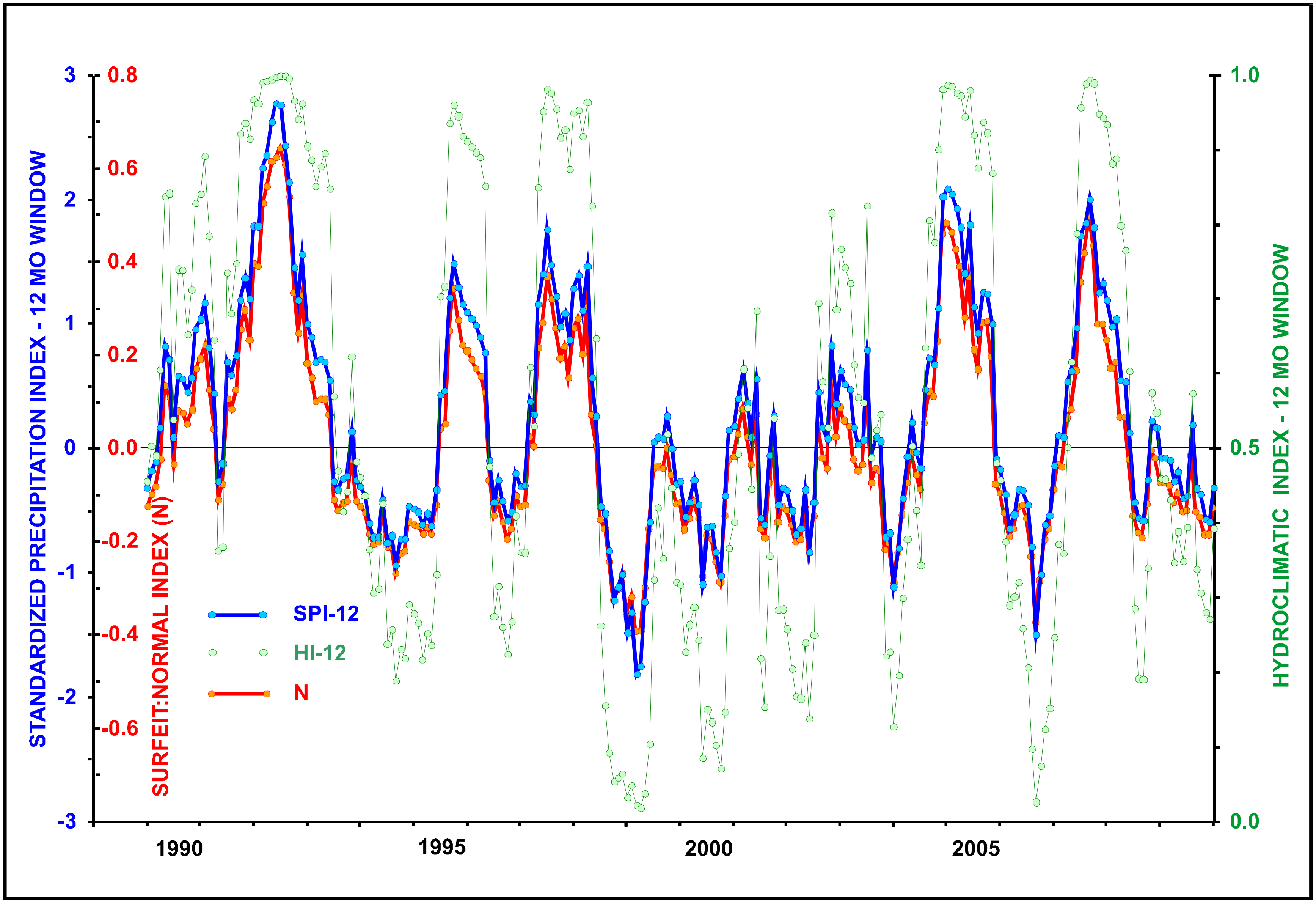

Figure 52 - Brazos River near Seymour moderate (1 year) memory indices (see Table 17), for 1990-2009

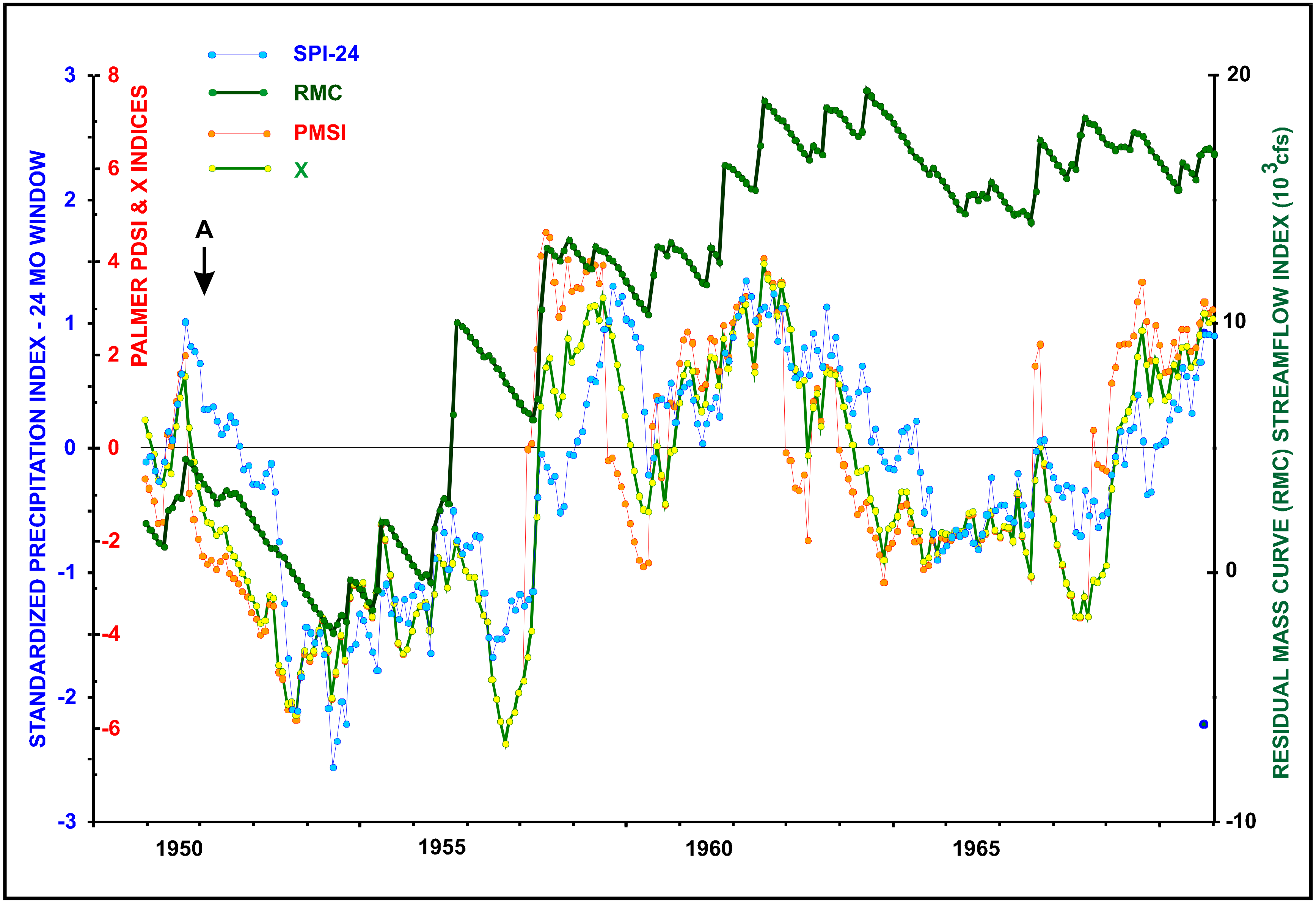

Figure 53 - Brazos River near Seymour long-term memory indices (see Table 17), for 1950-1969

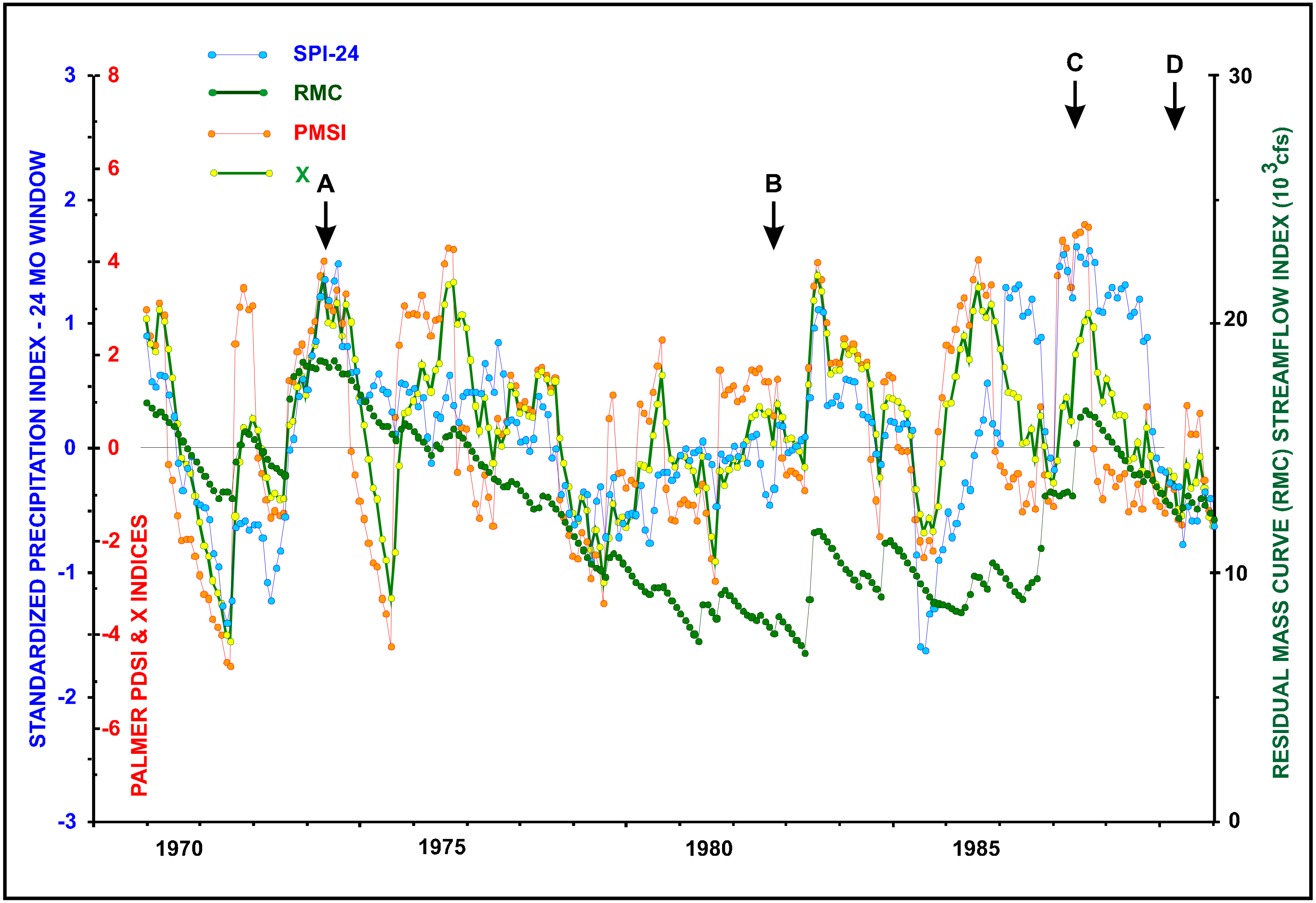

Figure 54 - Brazos River near Seymour long-term memory indices (see Table 17), for 1970-1989

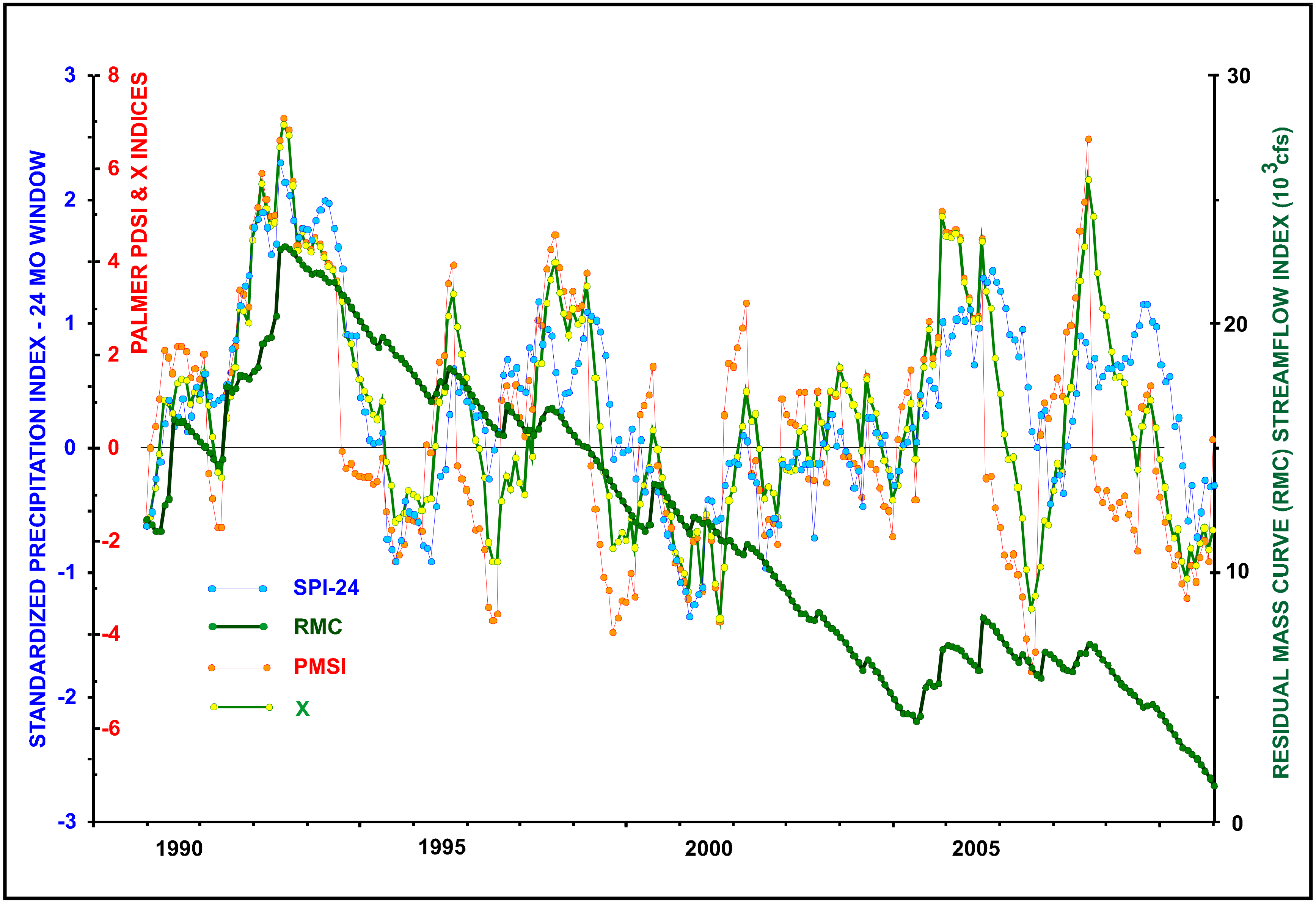

Figure 55 - Brazos River near Seymour long-term memory indices (see Table 17), for 1990-2009

4.6 San Bernard River at Boling

The San Bernard is a small river west of Houston. The drainage at the Boling gauge is about 1900 km^2, and lies entirely within the NCDC Upper Coast climatic division (Division 8). There are no reservoirs of any consequence upstream from the gauge site. Unfortunately, the record on this gauge is shorter than desirable, staring in May 1954, and the entire 1996 water year is missing. Nevertheless, this site is included here because it is the only river predominantly affected by the upper coastal plain environment. In Figs. 62-64, the range of the ordinate for RMC is 40K, and the origin varies. Table 22 summarizes the pairwise correlations among the various indices.

Table 22
Linear correlations between monthly index variables for San Bernard River at Boling
(Correlations greater than 0.7 in boldface)

	P	SPI-6	HI-6	SPI-12	HI-12	N	SPI-24	PDSI	X	RMC	Q
P	1.00	0.42	0.16	0.30	0.29	0.31	0.21	0.51	0.35	0.01	0.62
SPI-6		1.00	0.66	**0.75**	**0.70**	**0.73**	0.50	**0.84**	**0.78**	0.07	0.43
HI-6			1.00	0.55	0.54	0.55	0.37	0.59	0.58	0.02	0.31
SPI-12				1.00	**0.95**	**0.97**	**0.72**	**0.78**	**0.90**	0.14	0.34
HI-12					1.00	**0.98**	**0.70**	**0.75**	**0.92**	0.09	0.32
N						1.00	**0.70**	**0.76**	**0.91**	0.16	0.34
SPI-24							1.00	0.59	**0.82**	0.28	0.28
PDSI								1.00	**0.84**	-0.03	0.50
X									1.00	0.11	0.40
RMC										1.00	0.04
Q											1.00

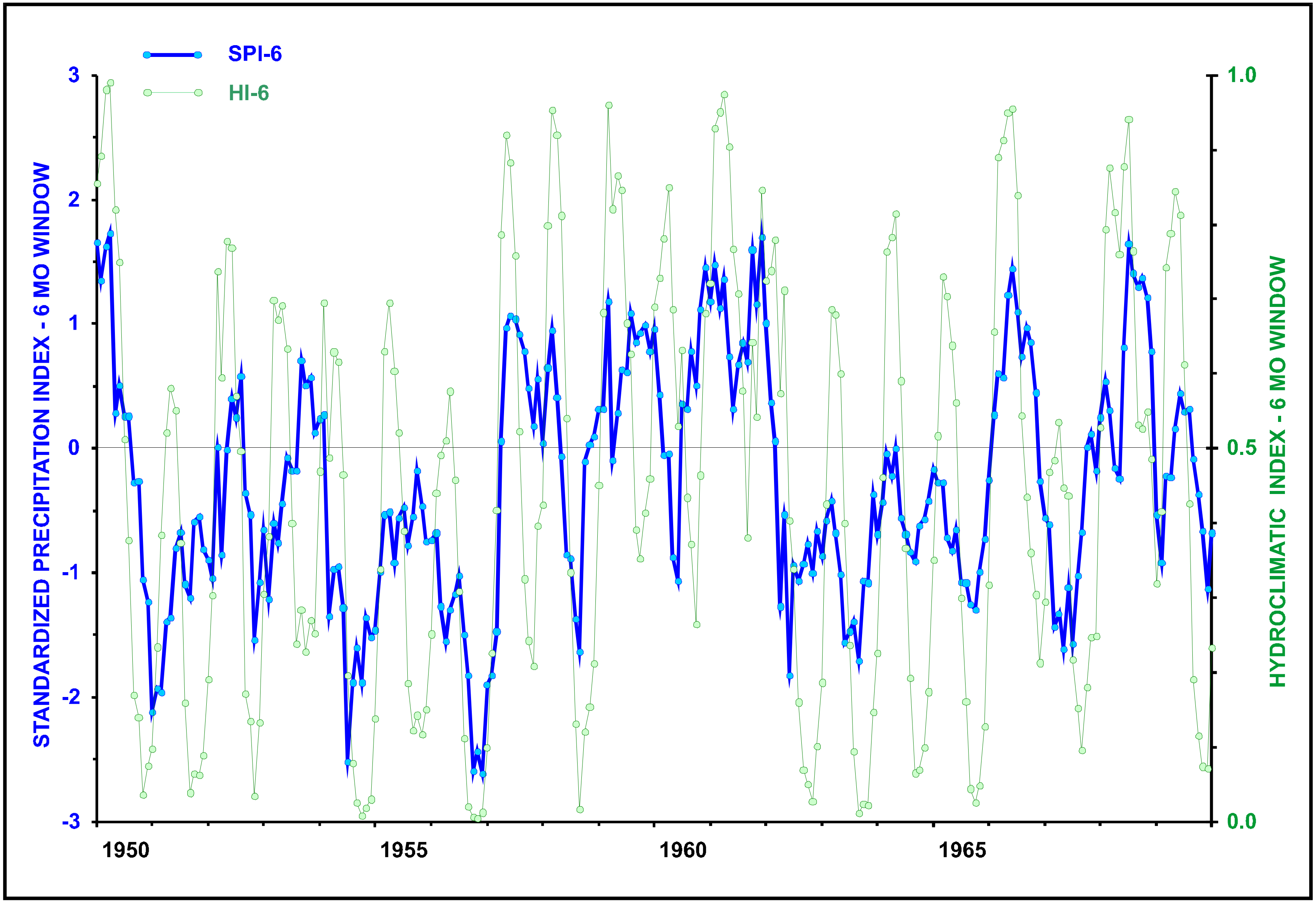

Figure 56 - San Bernard River at Boling short-term memory indices (see Table 17), for 1950-1969

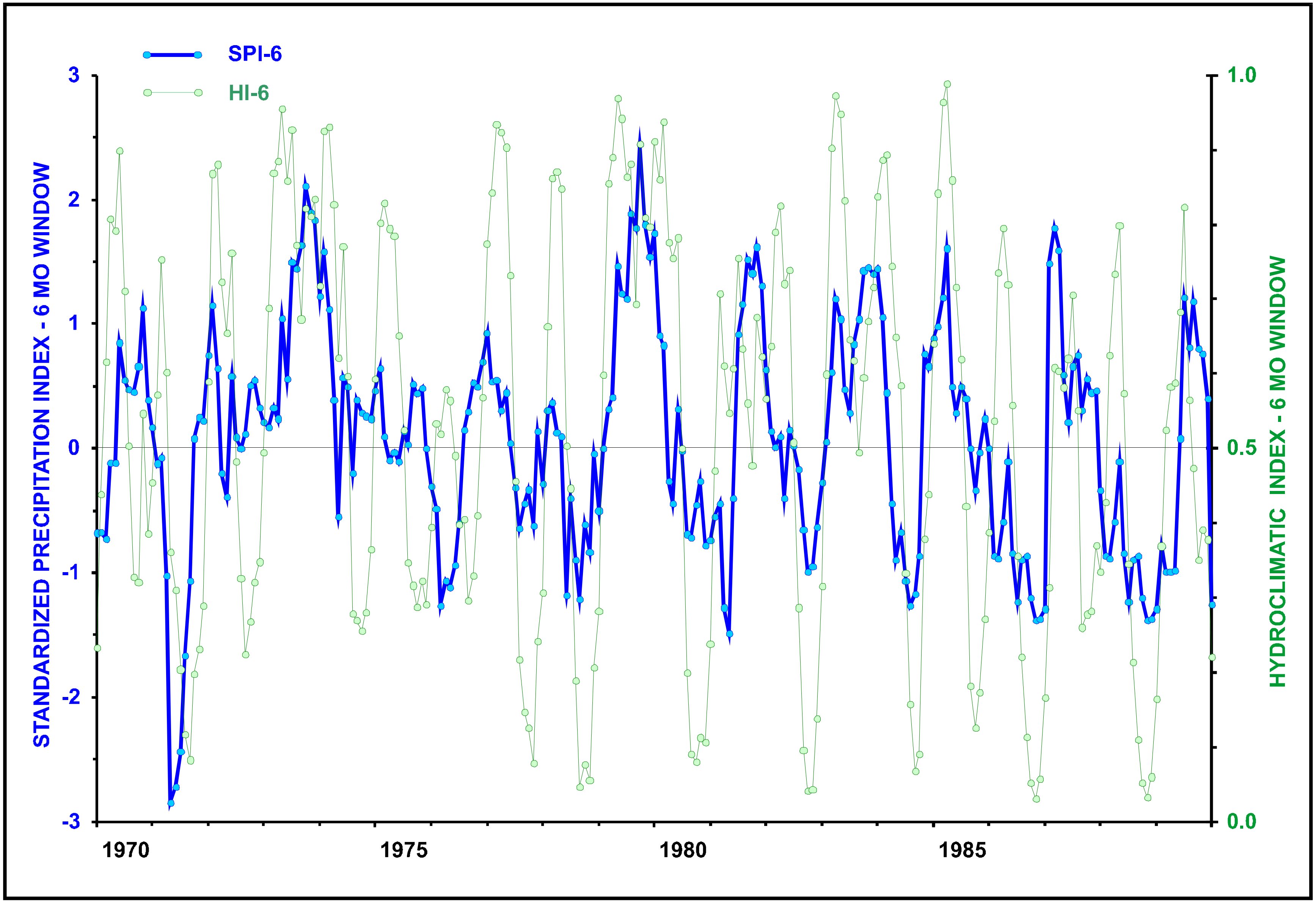

Figure 57 - San Bernard River at Boling short-term memory indices (see Table 17), for 1970-1989

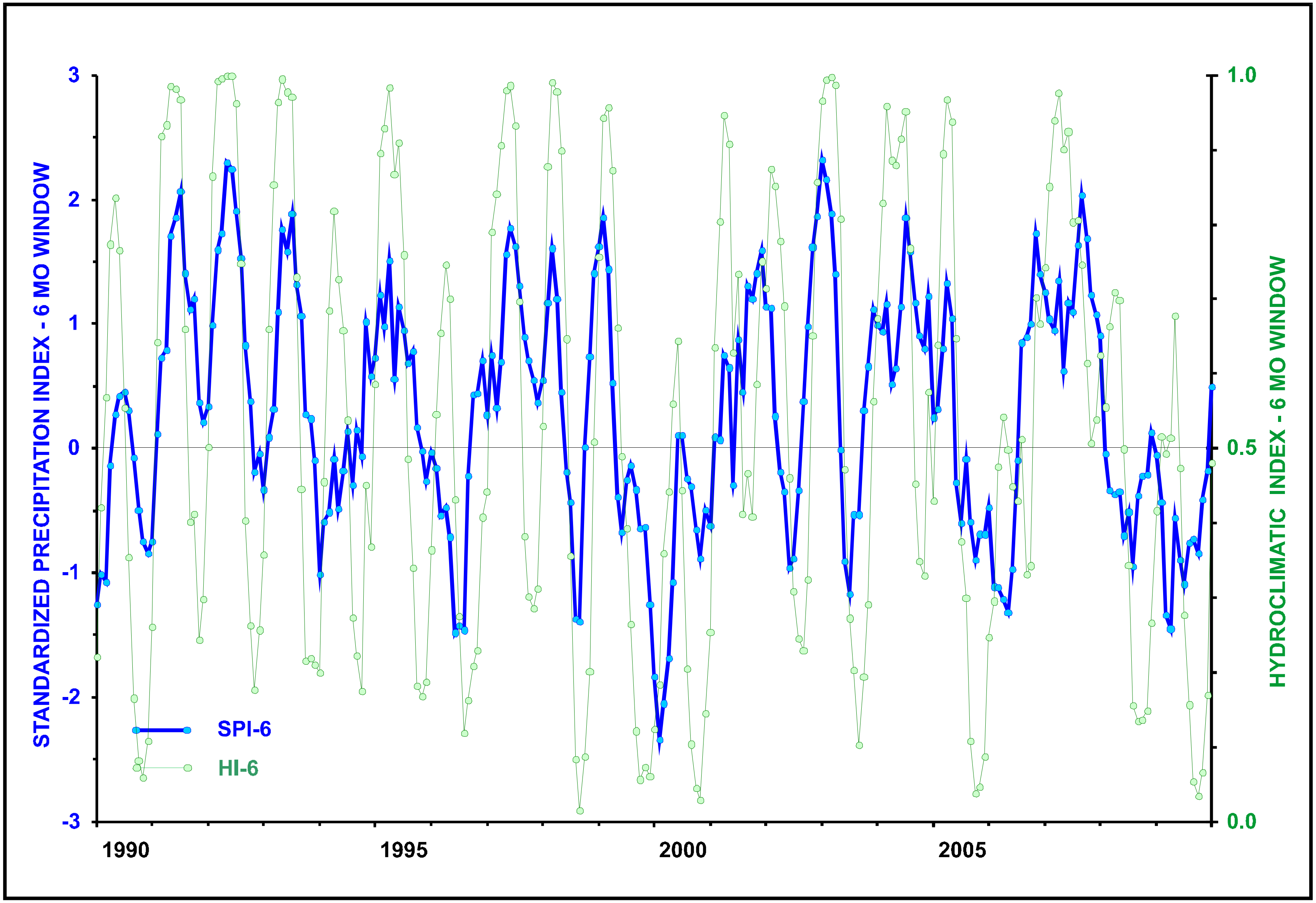

Figure 58 - San Bernard River at Boling short-term memory indices (see Table 17), for 1990-2009

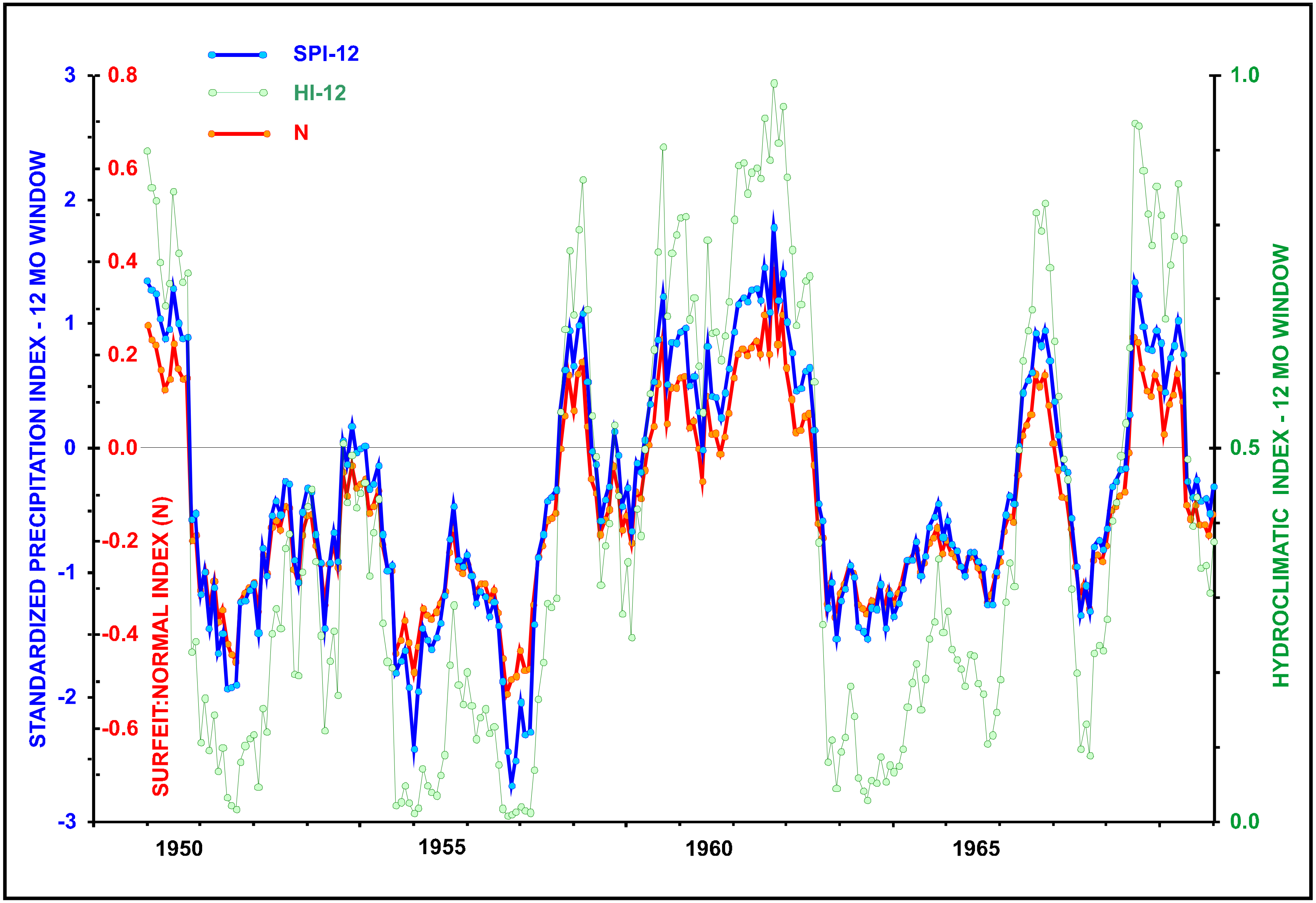

Figure 59 - San Bernard River at Boling moderate (1 year) memory indices (see Table 17), for 1950-1969

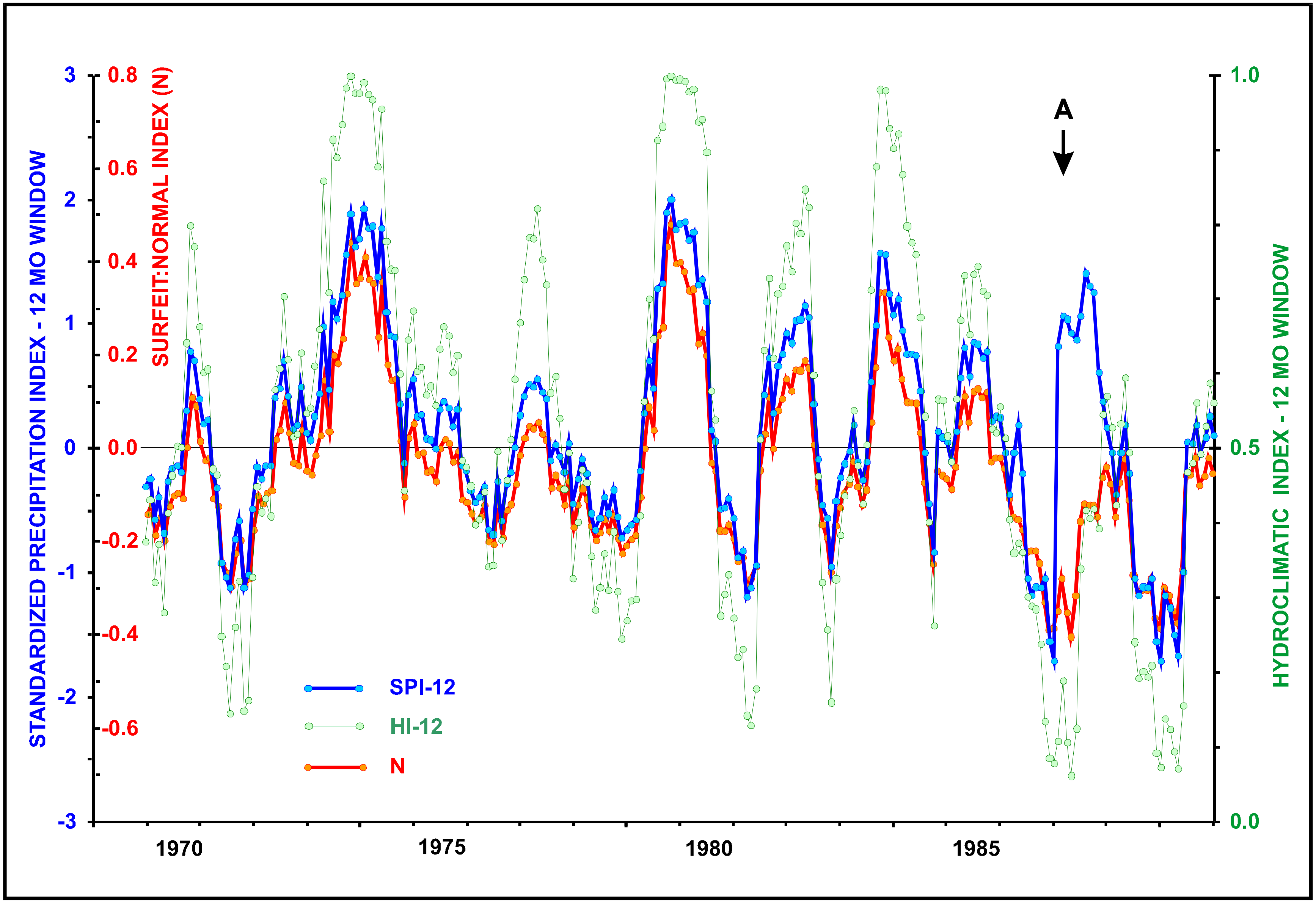

Figure 60 - San Bernard River at Boling moderate (1 year) memory indices (see Table 17), for 1970-1989

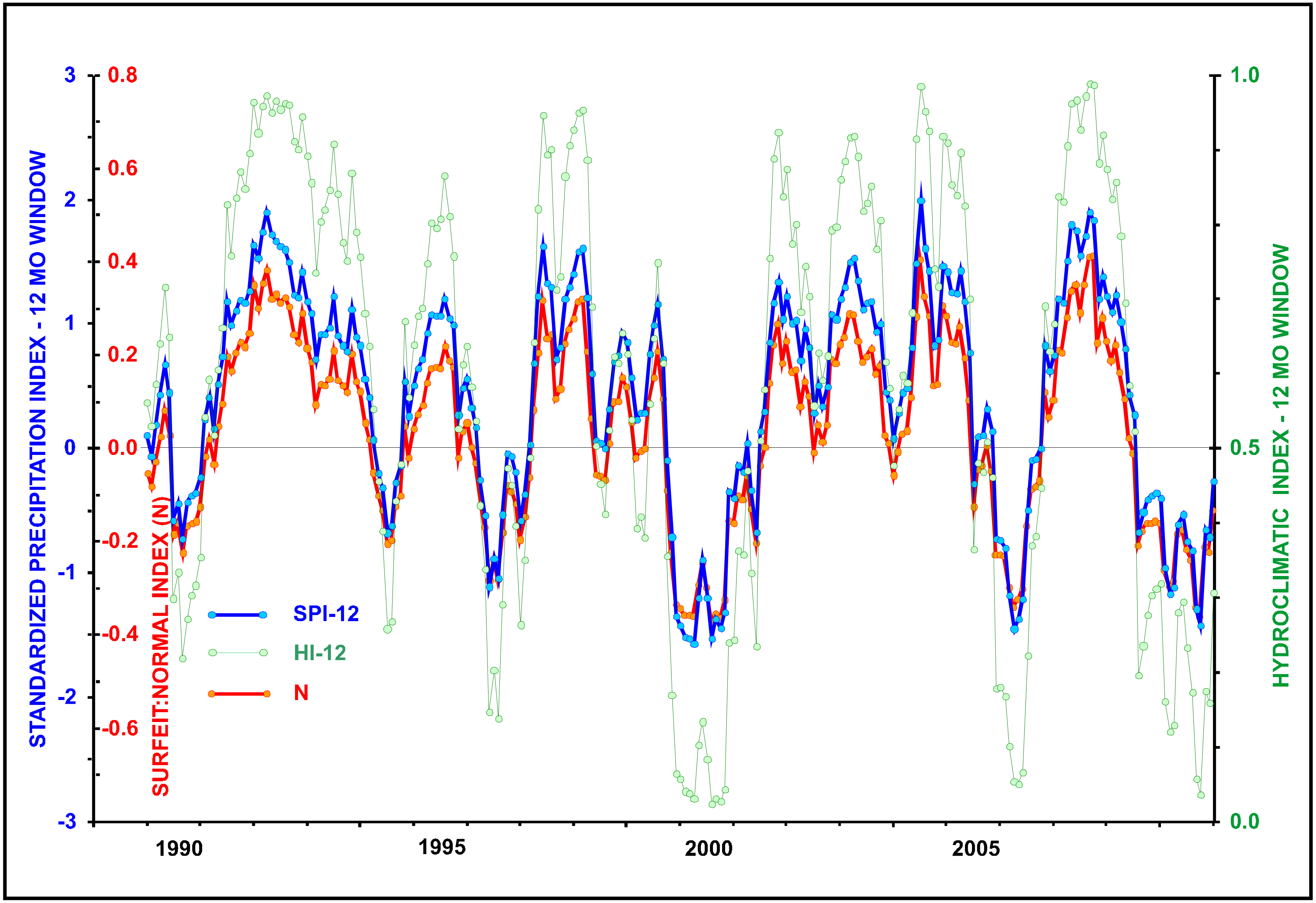

Figure 61 - San Bernard River at Boling moderate (1 year) memory indices (see Table 17), for 1990-2009

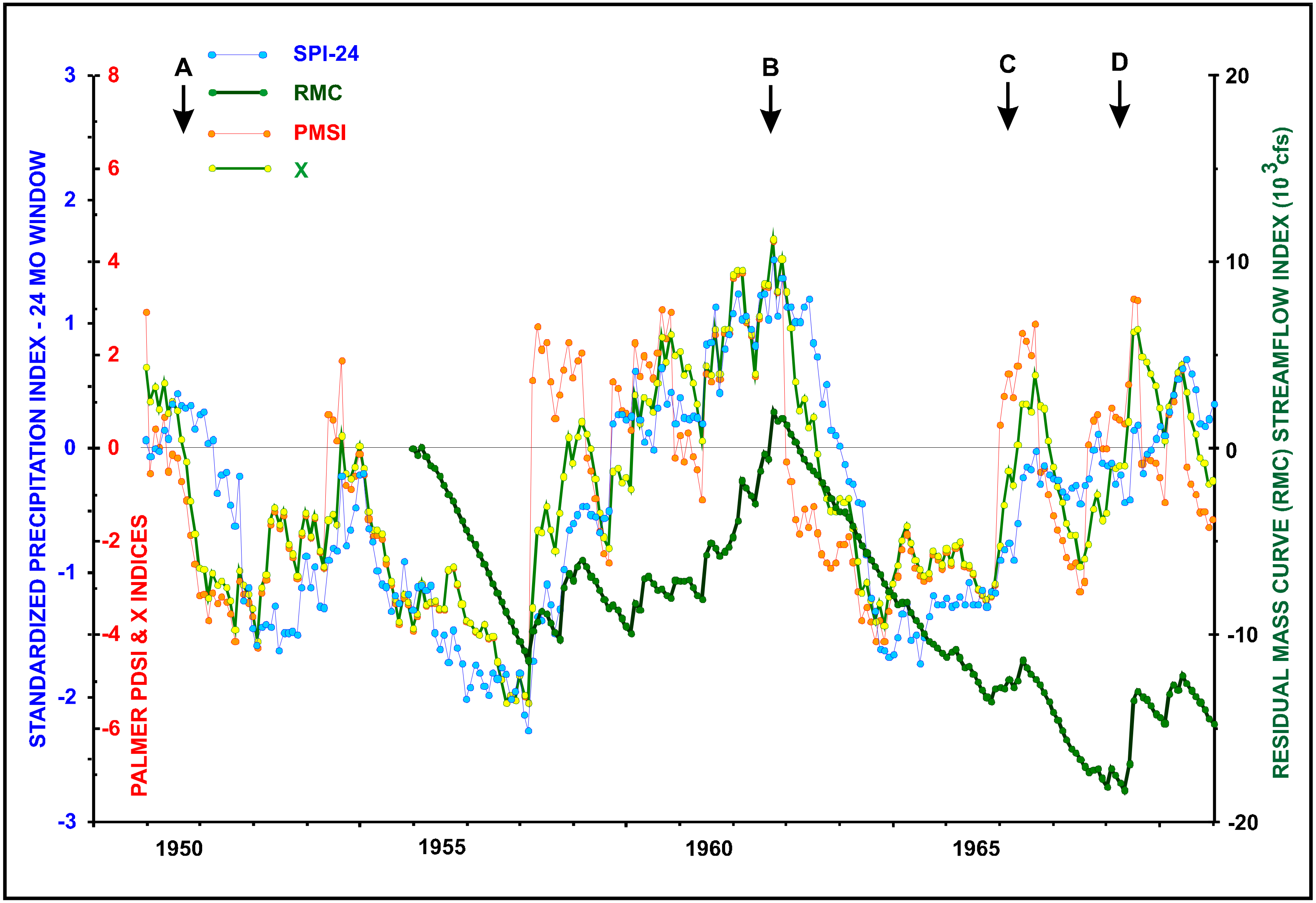

Figure 62 - San Bernard River at Boling long-term memory indices (see Table 17), for 1950-1969

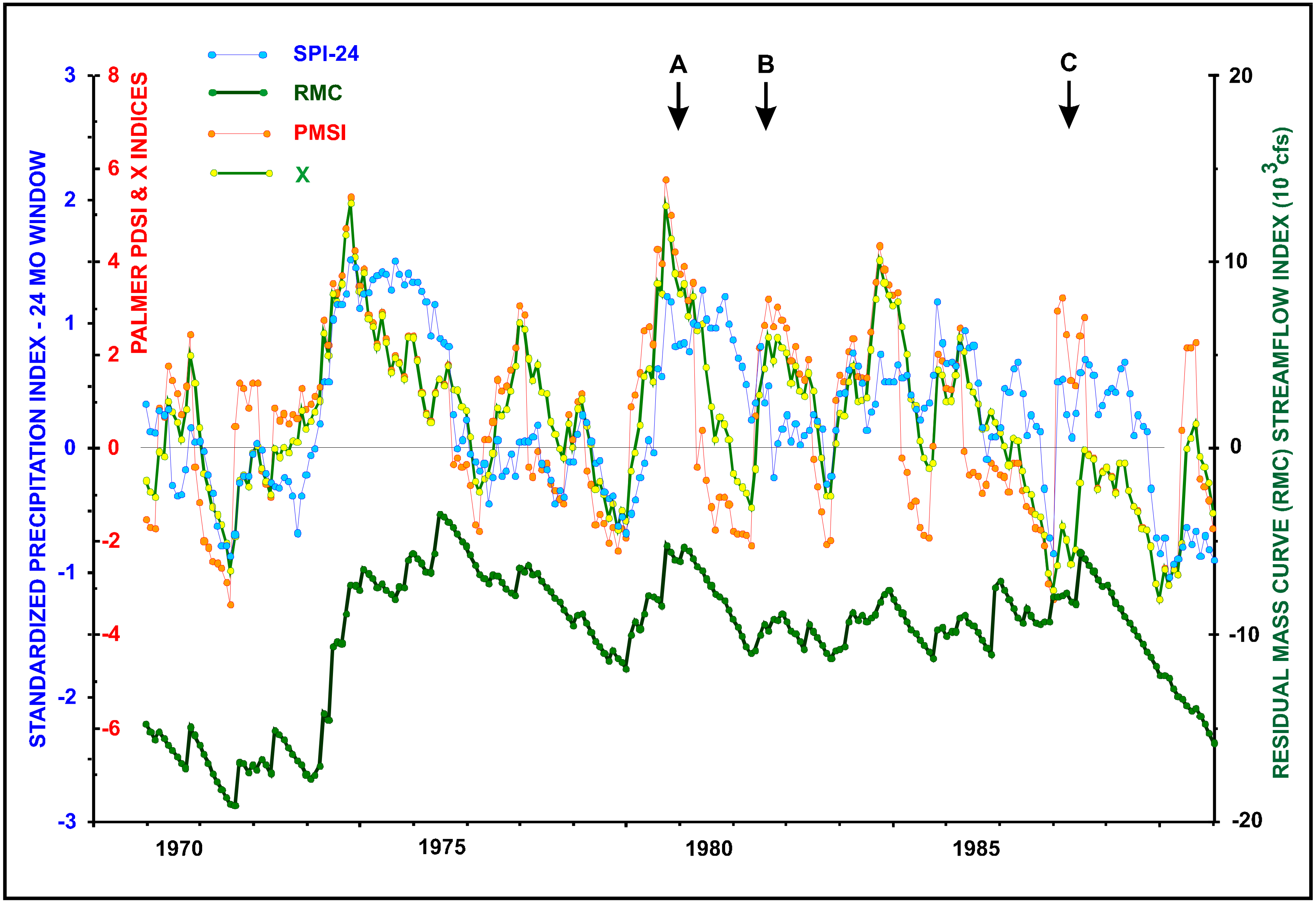

Figure 63 - San Bernard River at Boling long-term memory indices (see Table 17), for 1970-1989

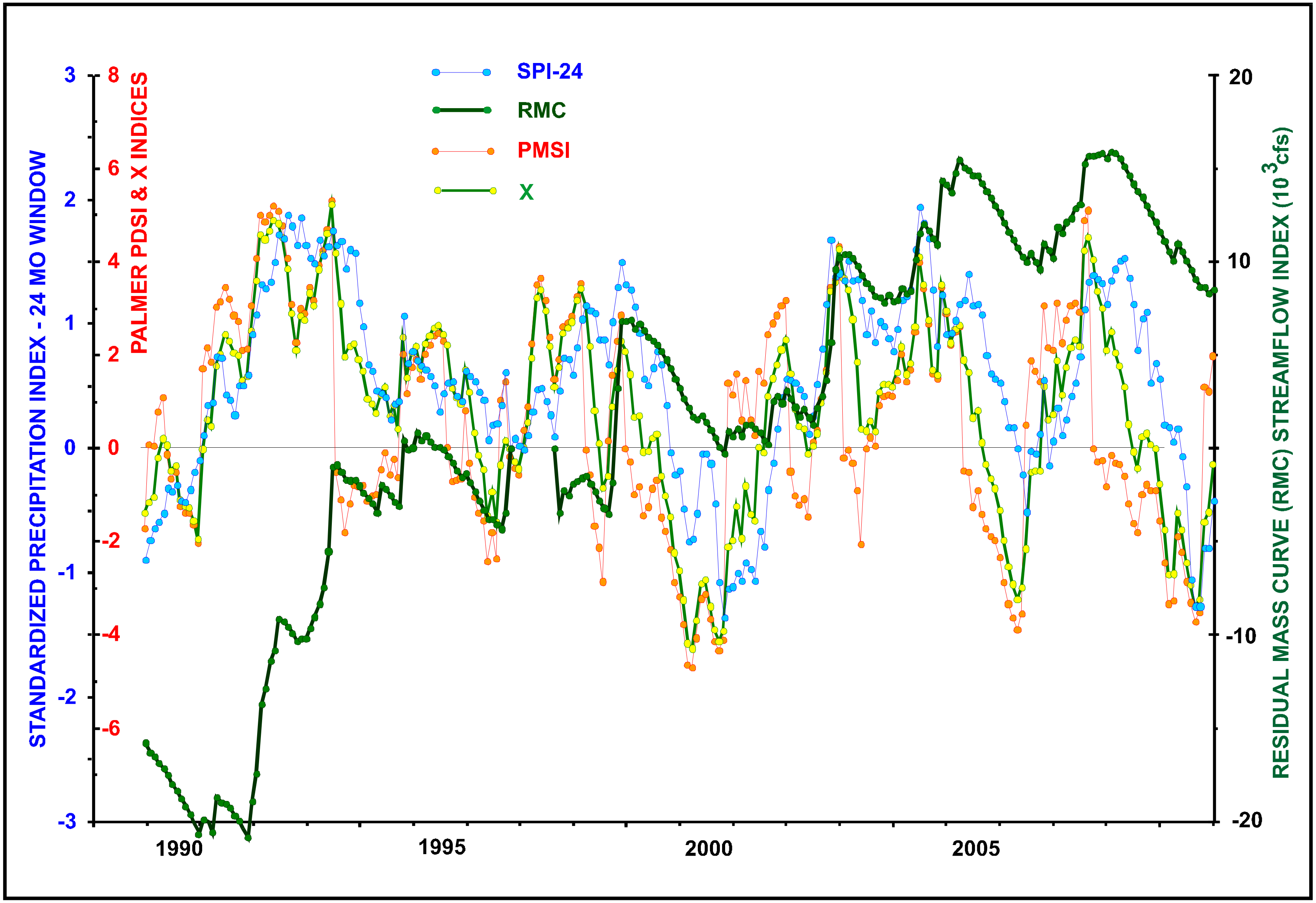

Figure 64 - San Bernard River at Boling long-term memory indices (see Table 17), for 1990-2009

4.7 Guadalupe River at Victoria

The Guadalupe River above the Victoria gauge is located primarily within the South Central climatic division (Division 7). There is one major reservoir on the main stem, Lake Canyon, a substantial distance upstream on the Balcones Escarpment thus affecting only the portion of the drainage from the Edwards Plateau. There is also a power-plant cooling lake on Coleto Creek. Total drainage area above the Victoria gauge is about 13,500 km^2, a little less than the area above the Seymour gauge on the Brazos. The gauge record is continuous since November 1934. The Guadalupe receives appreciable spring flow, generally maintaining its base flow. The range of the RMC plot, Figs. 71-73, is 200K (note the change of units on the y-axis), with a floating origin. Table 23 summarizes the pairwise correlations among the various indices.

Table 23
Linear correlations between monthly index variables for Guadalupe River at Victoria
(Correlations greater than 0.7 in boldface)

	P	SPI-6	HI-6	SPI-12	HI-12	N	SPI-24	PDSI	X	RMC	Q
P	1.00	0.42	0.15	0.31	0.28	0.31	0.18	0.46	0.30	-0.04	0.57
SPI-6		1.00	0.54	**0.76**	**0.70**	**0.74**	0.46	**0.84**	**0.75**	-0.04	0.54
HI-6			1.00	0.46	0.44	0.46	0.31	0.50	0.47	-0.05	0.37
SPI-12				1.00	**0.96**	**0.98**	**0.70**	**0.83**	**0.90**	0.01	0.48
HI-12					1.00	**0.97**	**0.71**	**0.80**	**0.92**	-0.03	0.44
N						1.00	0.68	**0.81**	**0.91**	0.02	0.47
SPI-24							1.00	0.63	**0.85**	0.10	0.34
PDSI								1.00	**0.86**	-0.12	0.55
X									1.00	0.00	0.46
RMC										1.00	0.02
Q											1.00

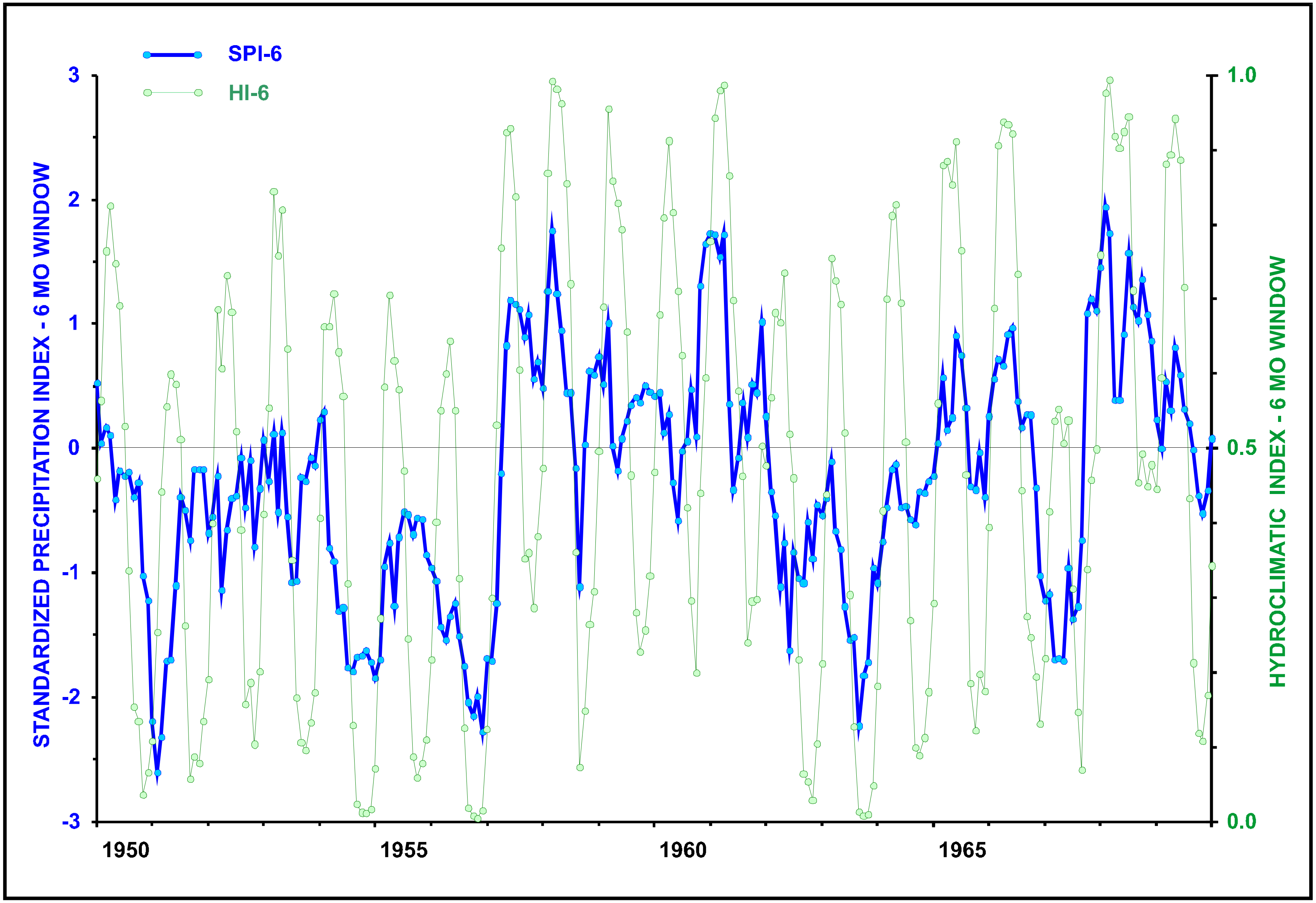

Figure 65 - Guadalupe River at Victoria short-term memory indices (see Table 17), for 1950-1969

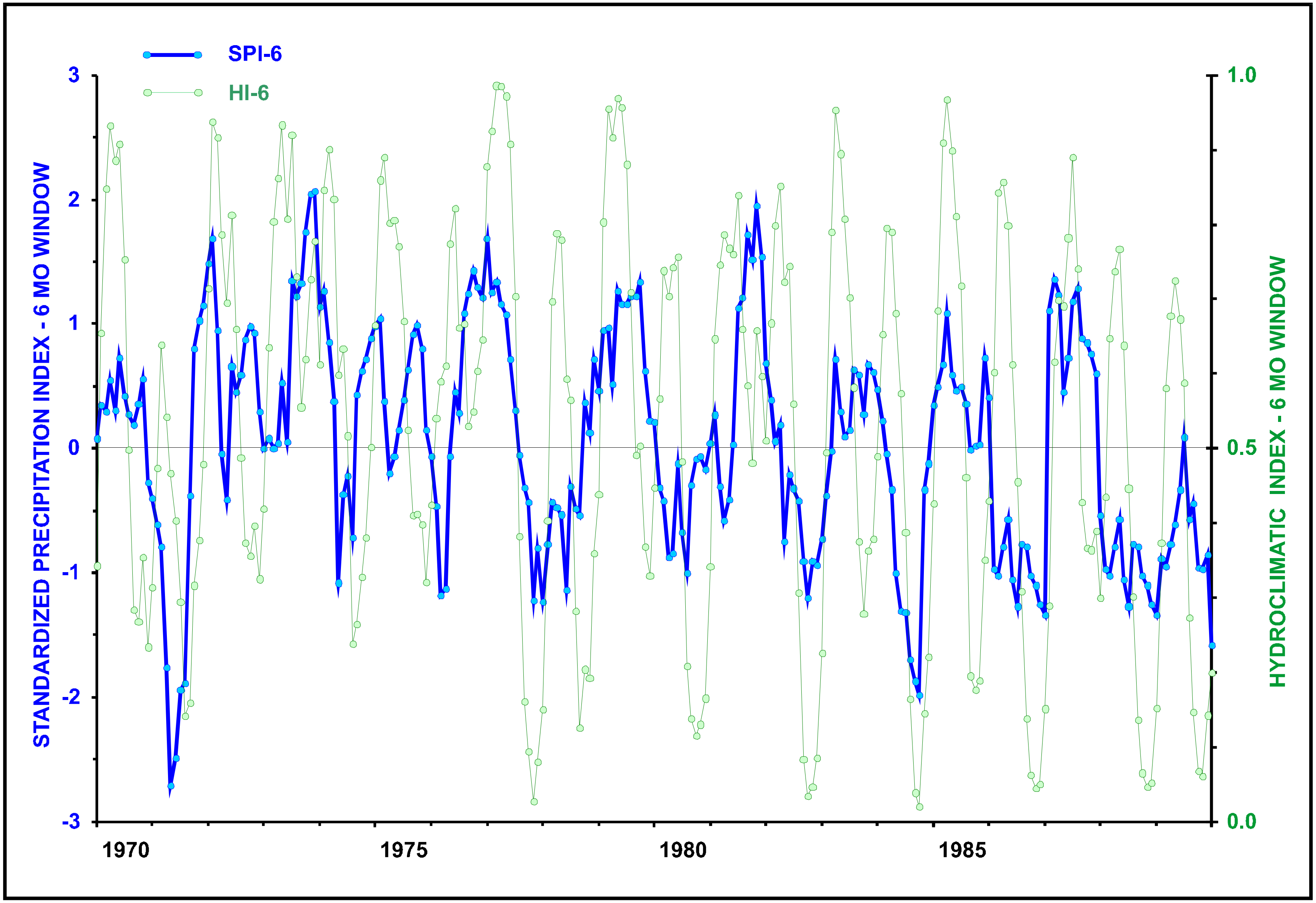

Figure 66 - Guadalupe River at Victoria short-term memory indices (see Table 17), for 1970-1989

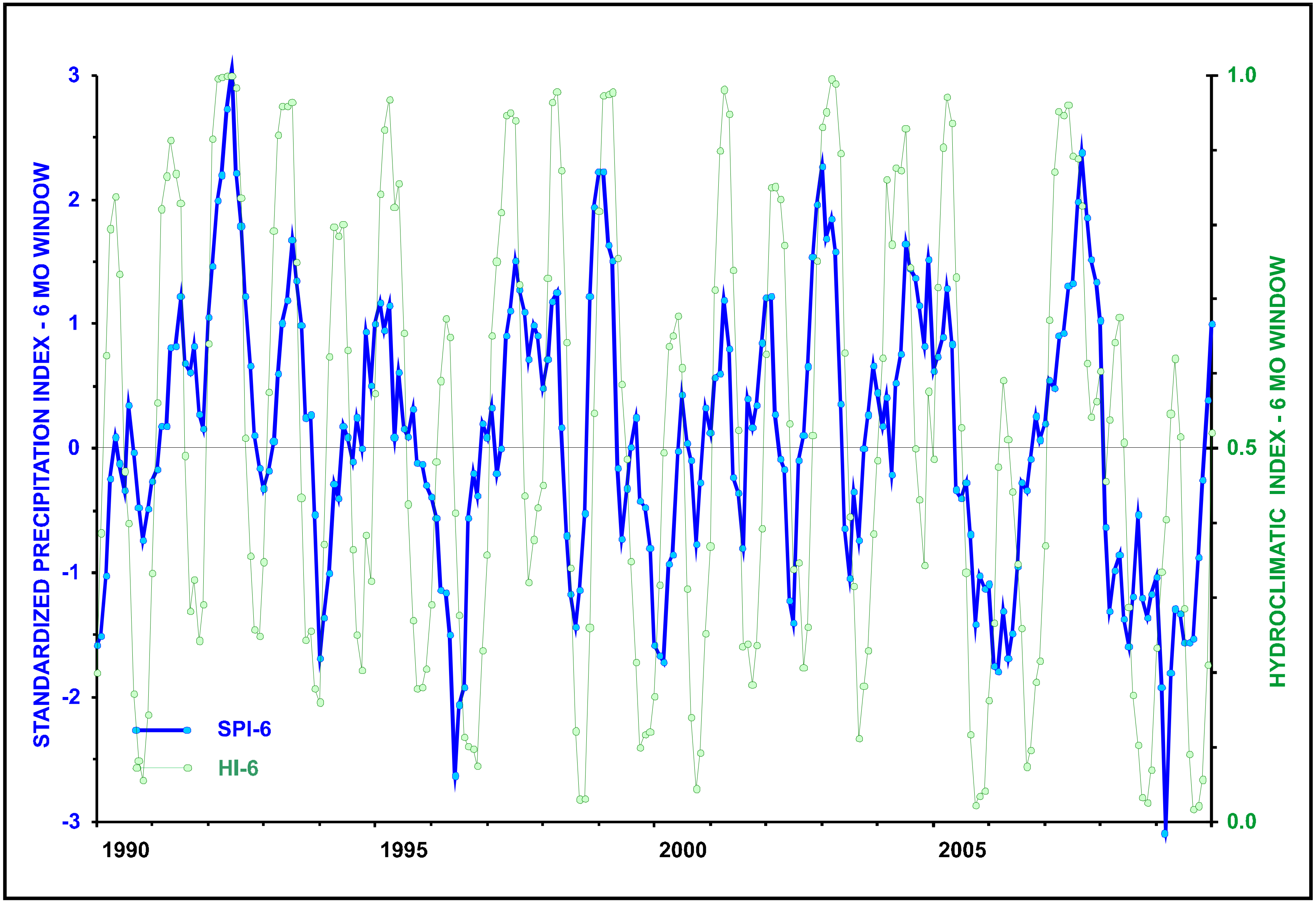

Figure 67 - Guadalupe River at Victoria short-term memory indices (see Table 17), for 1990-2009

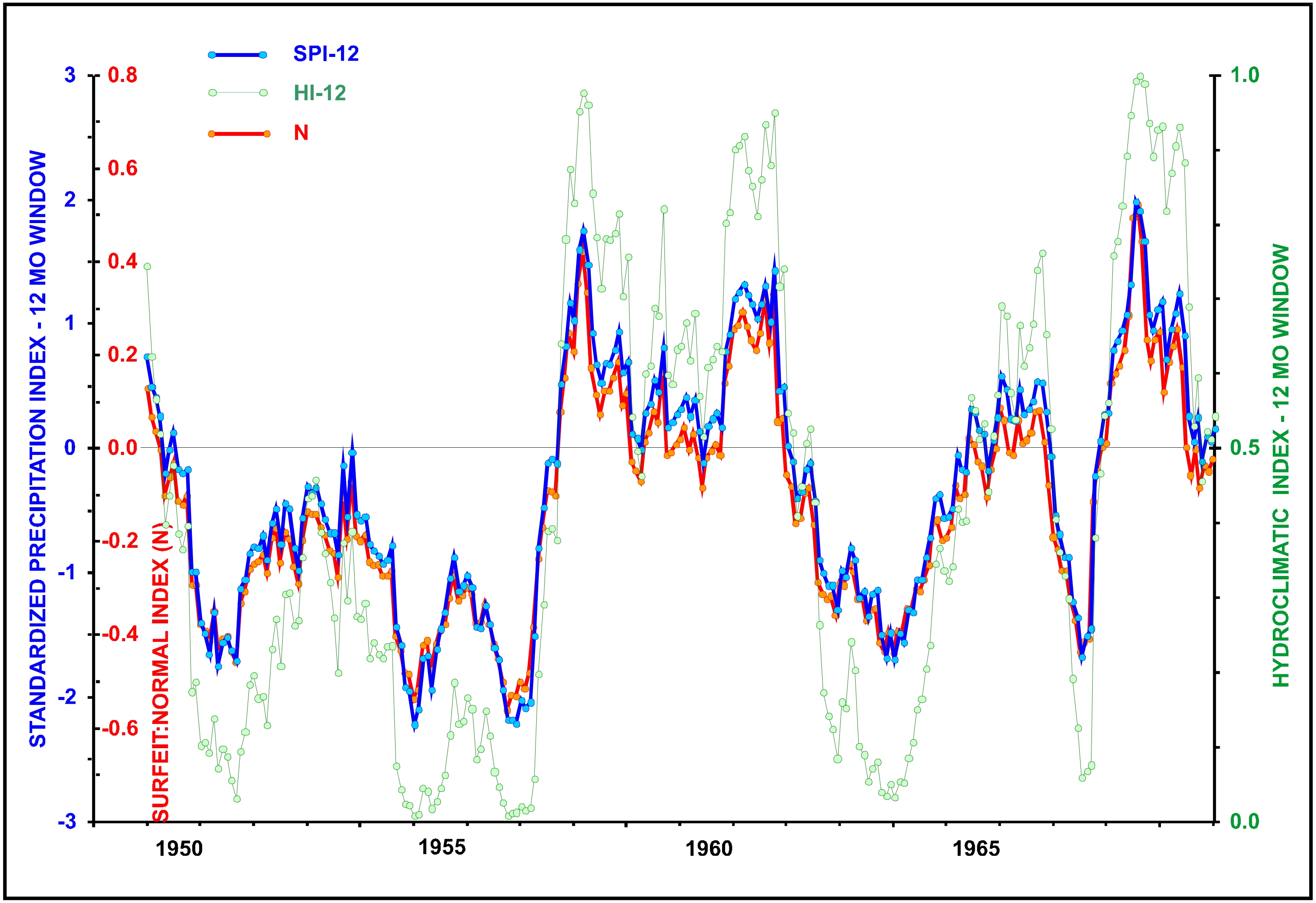

Figure 68 - Guadalupe River at Victoria moderate (1 year) memory indices (see Table 17), for 1950-1969

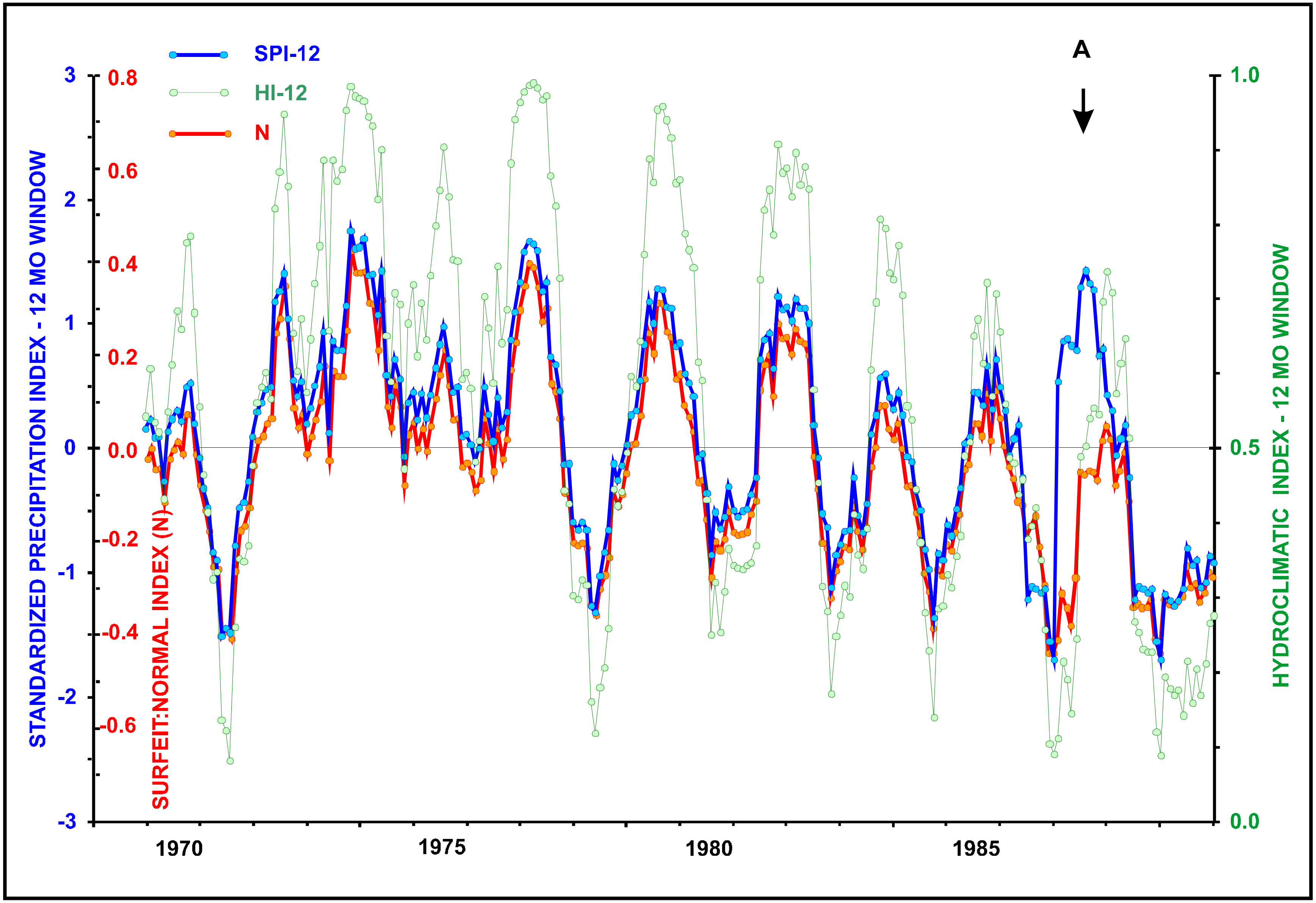

Figure 69 - Guadalupe River at Victoria moderate (1 year) memory indices (see Table 17), for 1970-1989

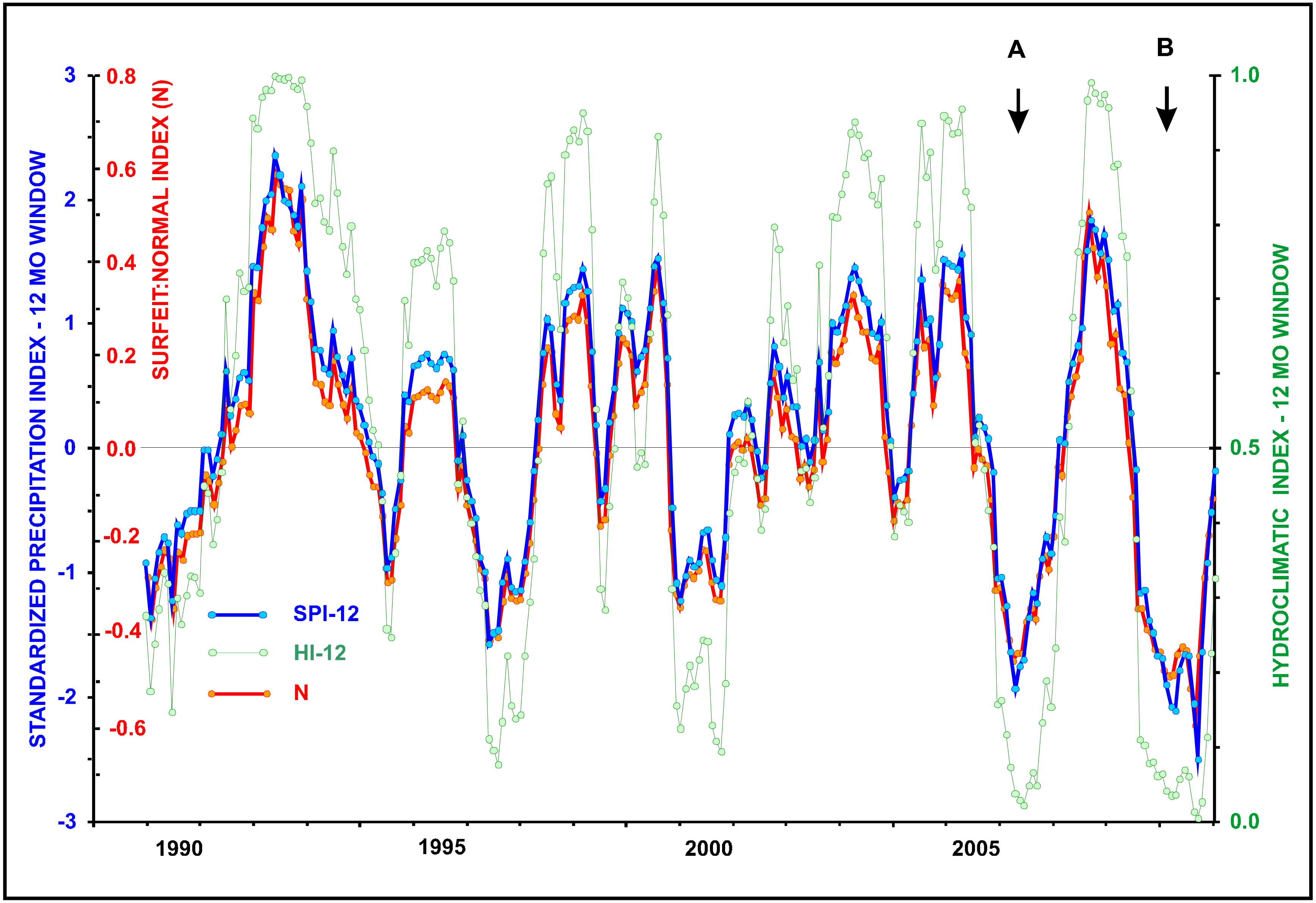

Figure 70 - Guadalupe River at Victoria moderate (1 year) memory indices (see Table 17), for 1990-2009

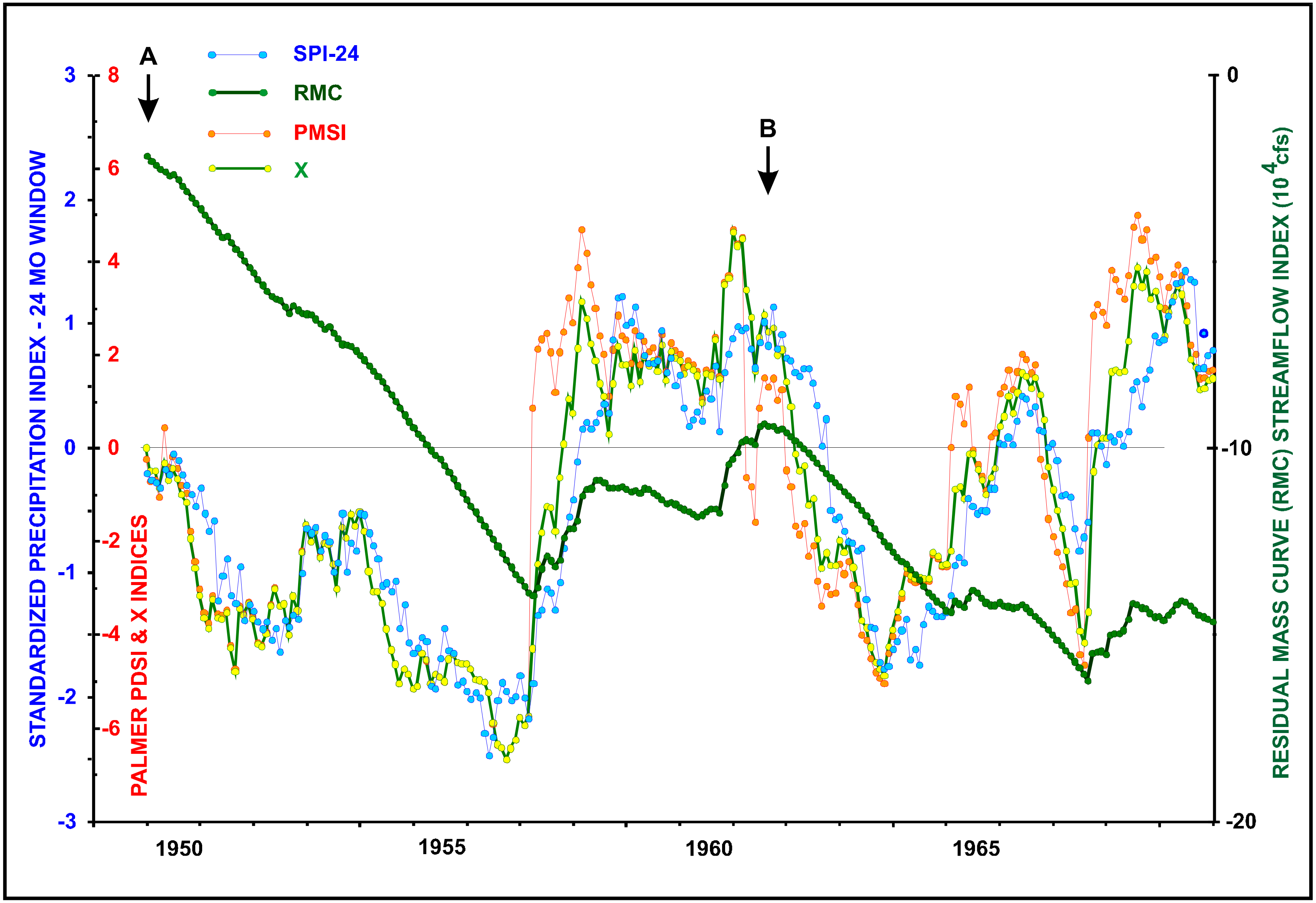

Figure 71 - Guadalupe River at Victoria long-term memory indices (see Table 17), for 1950-1969

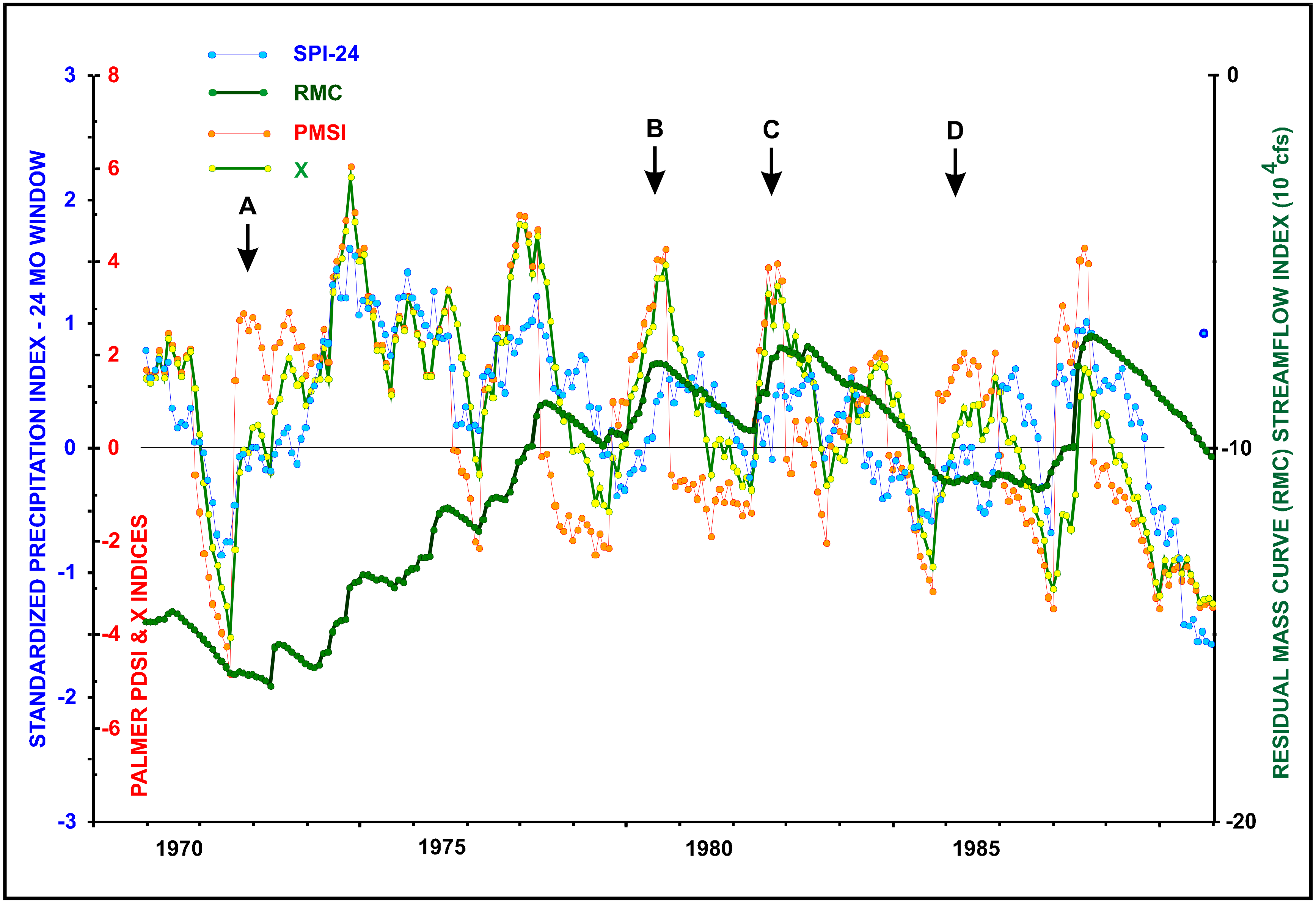

Figure 72 - Guadalupe River at Victoria long-term memory indices (see Table 17), for 1970-1989

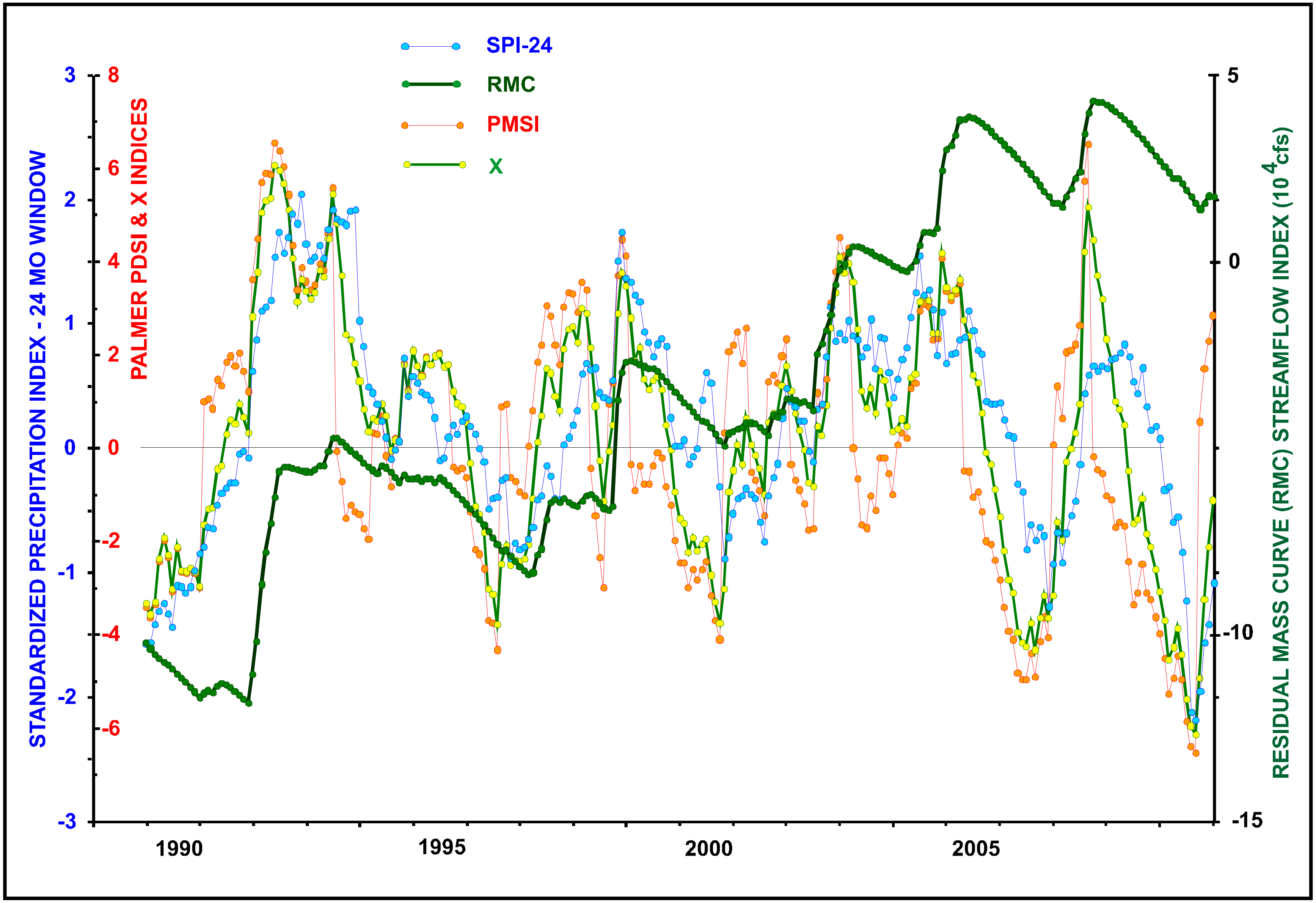

Figure 73 - Guadalupe River at Victoria long-term memory indices (see Table 17), for 1990-2009

4.8 San Antonio River at Goliad

Most of the drainage of the San Antonio lies within the South Central climatic division, except for the upper reaches of Medina River and Cibolo Creek. The Goliad gauge has a watershed of 40,000 km^2, in which there are three major reservoirs, two power plant lakes near San Antonio, and one water-supply reservoir in the upper reach of Medina Creek. While flow records date back to 1924, there are major gaps, and the continuous record begins in March 1939. USGS has never logged a zero daily flow here.

Because the climatic division for this gauge is the same as the Guadalupe (Section 4.6, above), the climatological indices are the same as shown in Figs. 65-73, so there is no need to repeat them here. Because the RMC index is different, only the indices with long-term memory are plotted in Figures 74-76. The range on the RMC plot is 90K (note change of axis unit) with a floating origin. Table 24 summarizes the pairwise correlations among the various indices, which are identical to those in Table 23 except for those involving either the RMC or Q.

Table 24
Linear correlations between monthly index variables for San Antonio River at Goliad
(Correlations greater than 0.7 in boldface)

	P	SPI-6	HI-6	SPI-12	HI-12	N	SPI-24	PDSI	X	RMC	Q
P	1.00	0.42	0.15	0.31	0.28	0.31	0.18	0.46	0.30	-0.04	0.57
SPI-6		1.00	0.54	**0.76**	**0.70**	**0.74**	0.46	**0.84**	**0.75**	-0.07	0.43
HI-6			1.00	0.46	0.44	0.46	0.31	0.50	0.47	-0.07	0.26
SPI-12				1.00	**0.96**	**0.98**	**0.70**	**0.83**	**0.90**	-0.04	0.37
HI-12					1.00	**0.97**	**0.71**	**0.80**	**0.92**	-0.08	0.33
N						1.00	0.68	**0.81**	**0.91**	-0.03	0.36
SPI-24							1.00	0.63	**0.85**	0.00	0.27
PDSI								1.00	**0.86**	-0.15	0.44
X									1.00	-0.06	0.36
RMC										1.00	0.02
Q											1.00

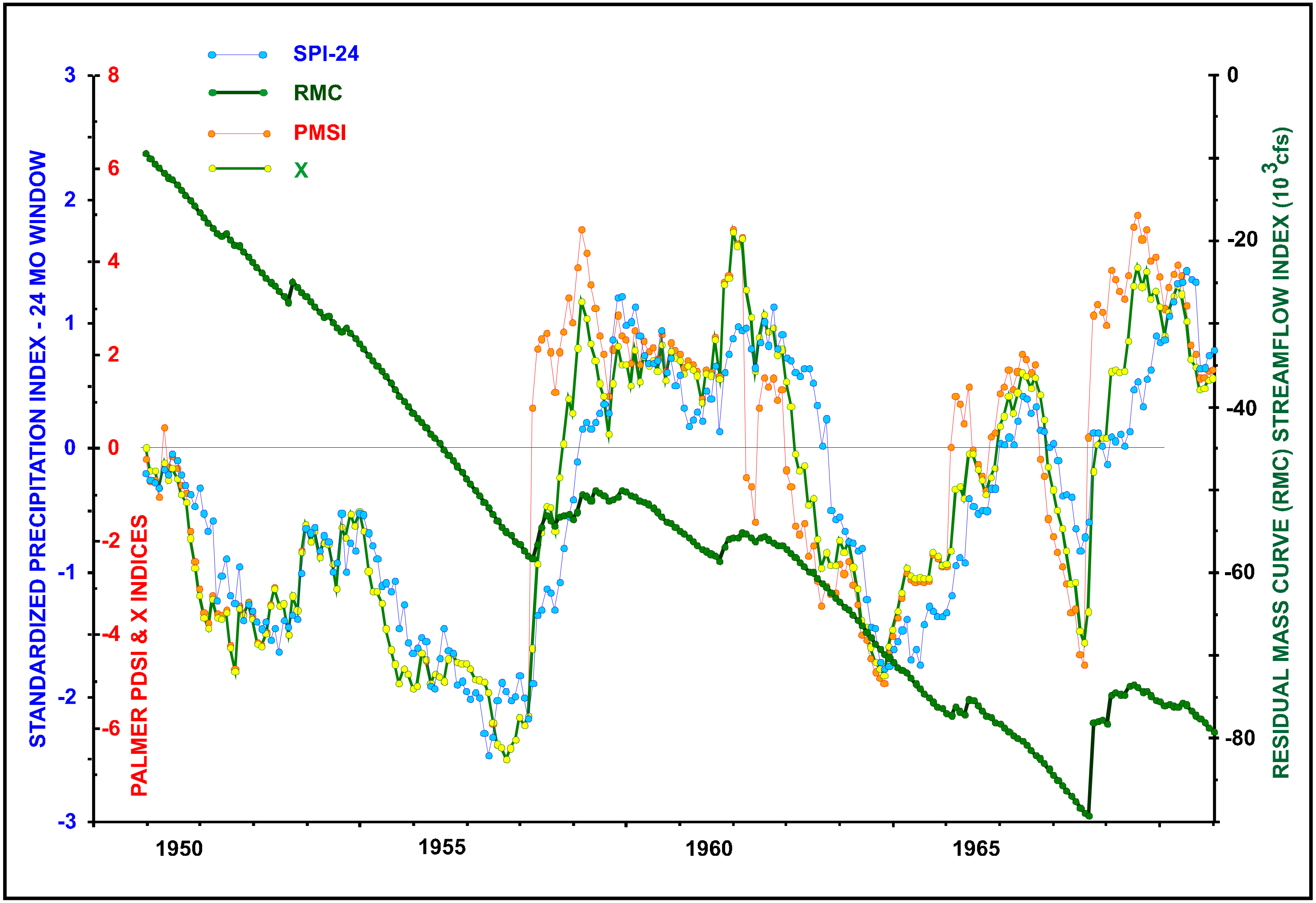

Figure 74 - San Antonio River at Goliad long-term memory indices (see Table 17), for 1950-1969

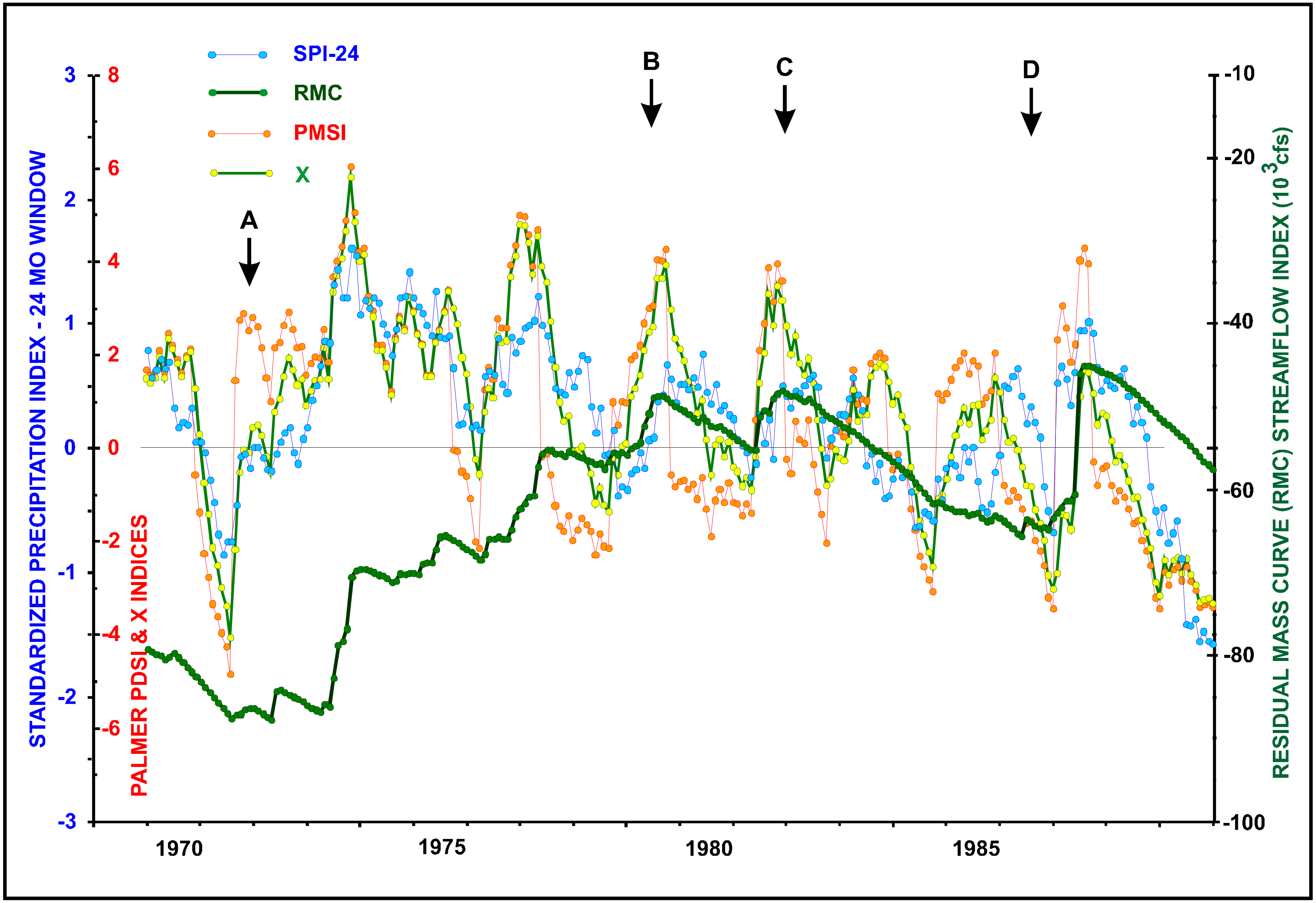

Figure 75 - San Antonio River at Goliad long-term memory indices (see Table 17), for 1970-1989

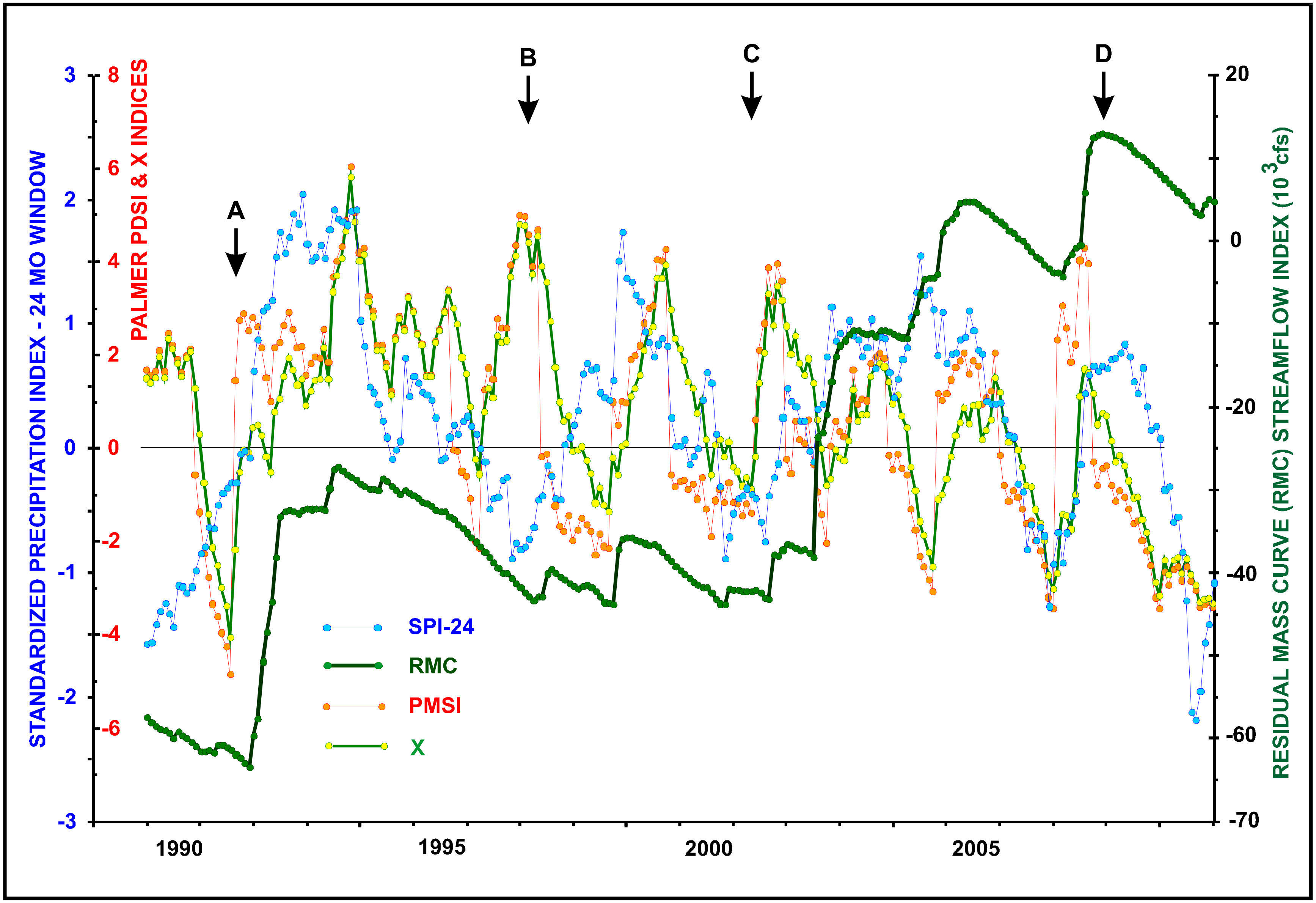

Figure 76 - San Antonio River at Goliad long-term memory indices (see Table 17), for 1990-2009

4.9 Nueces River at Cotulla

The selection of Cotulla instead of Three Rivers on the Nueces was made to avoid the impacts of Choke Canyon reservoir upstream from Three Rivers on the Frio. The Cotulla gauge watershed area is some 13,400 km^2, about one-third that of Three Rivers, but most of the difference between the two gauges is behind Choke Canyon dam. The Nueces rises in the Edwards Plateau climatic division (Division 6) and traverses much of the South Texas division (Division 9) to the gauge at Cotulla. Since about 65% of the Cotulla drainage is in the South Texas division, for expedience the South Texas climatology was assumed. The flow record at Cotulla extends back to October of 1926. The watershed is semi-arid, and zero daily flows occur frequently in the record. Of the 1927-2011 record of *monthly* flows, more than 25% are zeroes. The range on the RMC plot is 20K with a floating origin. Table 25 summarizes the pairwise correlations among the various indices.

Table 25
Linear correlations between monthly index variables for Nueces River at Cotulla
(Correlations greater than 0.7 in boldface)

	P	SPI-6	HI-6	SPI-12	HI-12	N	SPI-24	PDSI	X	RMC	Q
P	1.00	0.38	0.05	0.27	0.23	0.27	0.19	0.45	0.29	0.08	0.45
SPI-6		1.00	0.48	**0.74**	0.64	**0.71**	0.47	**0.80**	**0.71**	0.25	0.26
HI-6			1.00	0.41	0.40	0.41	0.30	0.44	0.42	0.12	0.04
SPI-12				1.00	**0.94**	**0.98**	**0.71**	**0.80**	**0.89**	0.37	0.19
HI-12					1.00	**0.94**	**0.75**	**0.76**	**0.90**	0.38	0.16
N						1.00	**0.70**	**0.80**	**0.90**	0.35	0.19
SPI-24							1.00	0.65	**0.85**	0.52	0.11
PDSI								1.00	**0.86**	0.34	0.28
X									1.00	0.45	0.19
RMC										1.00	0.05
Q											1.00

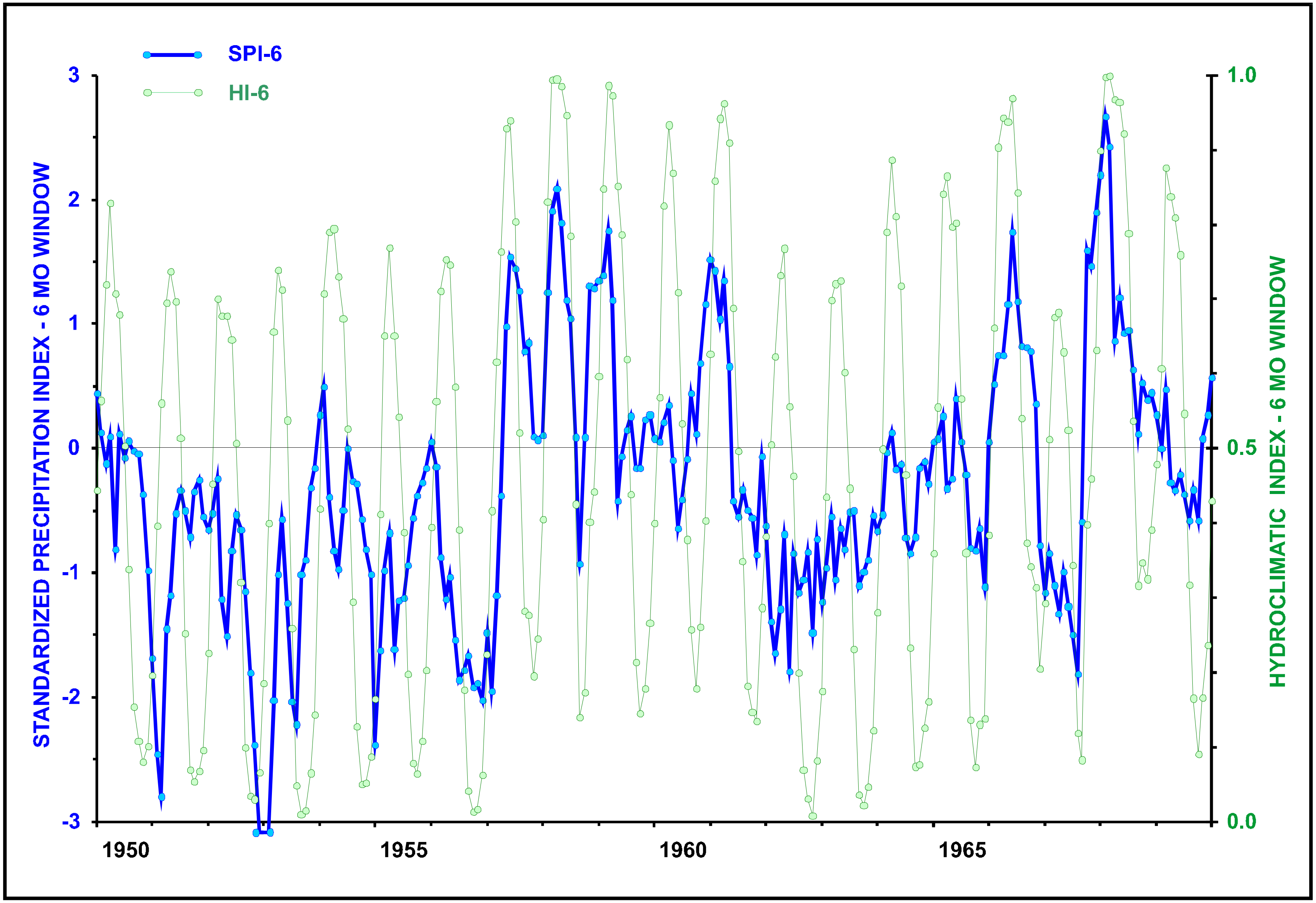

Figure 77 - Nueces River at Cotulla short-term memory indices (see Table 17), for 1950-1969

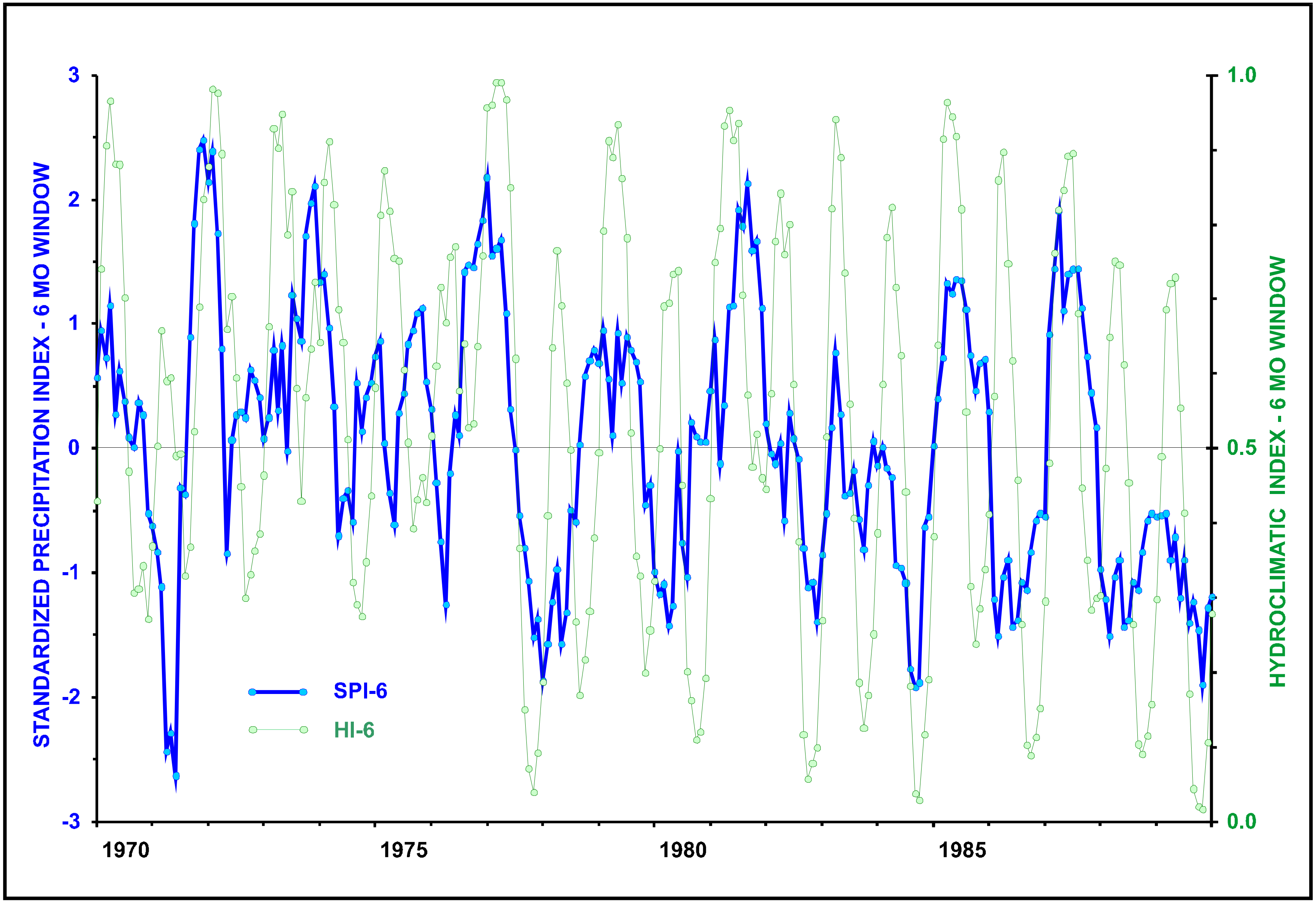

Figure 78 - Nueces River at Cotulla short-term memory indices (see Table 17), for 1970-1989

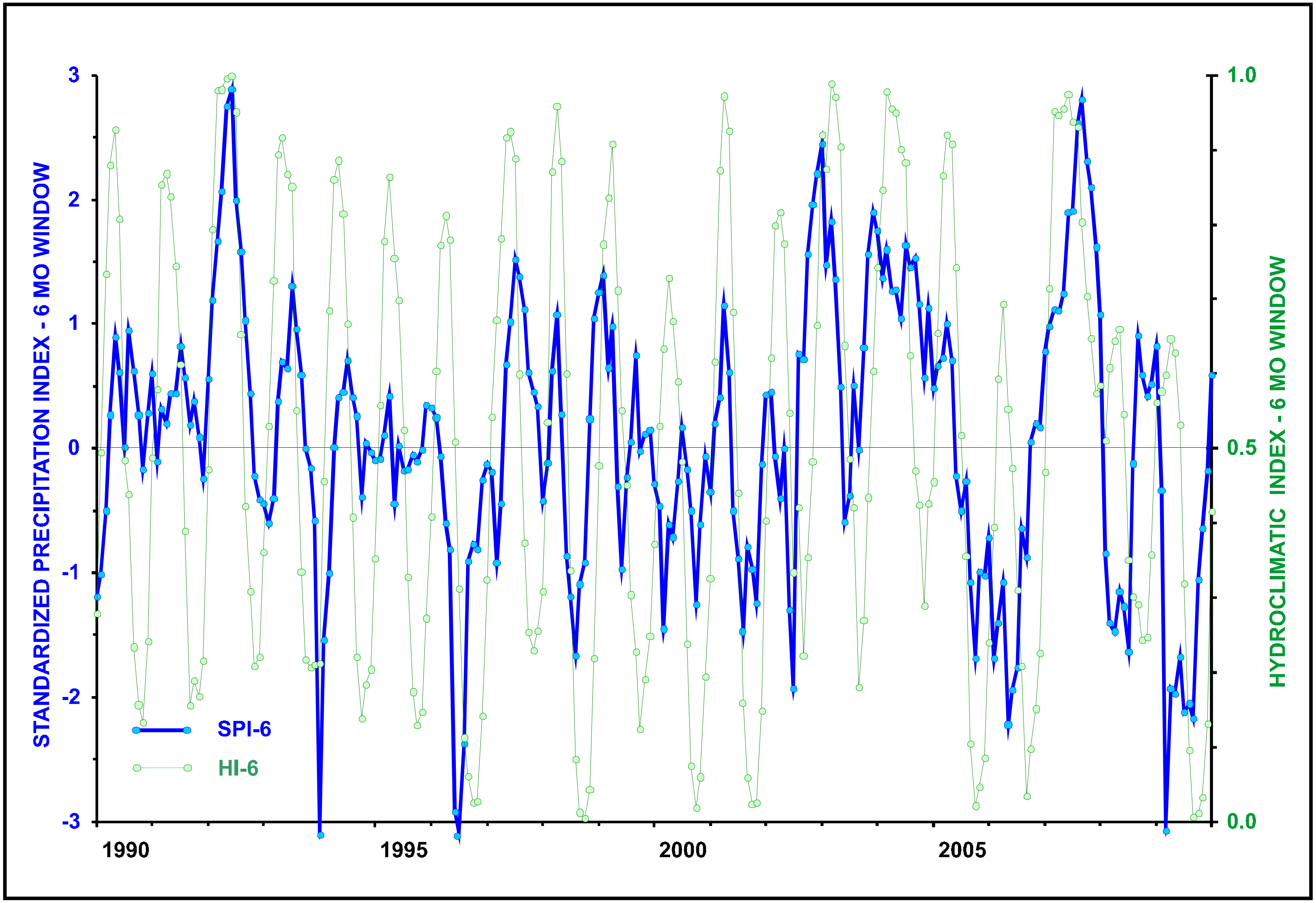

Figure 79 - Nueces River at Cotulla short-term memory indices (see Table 17), for 1990-2009

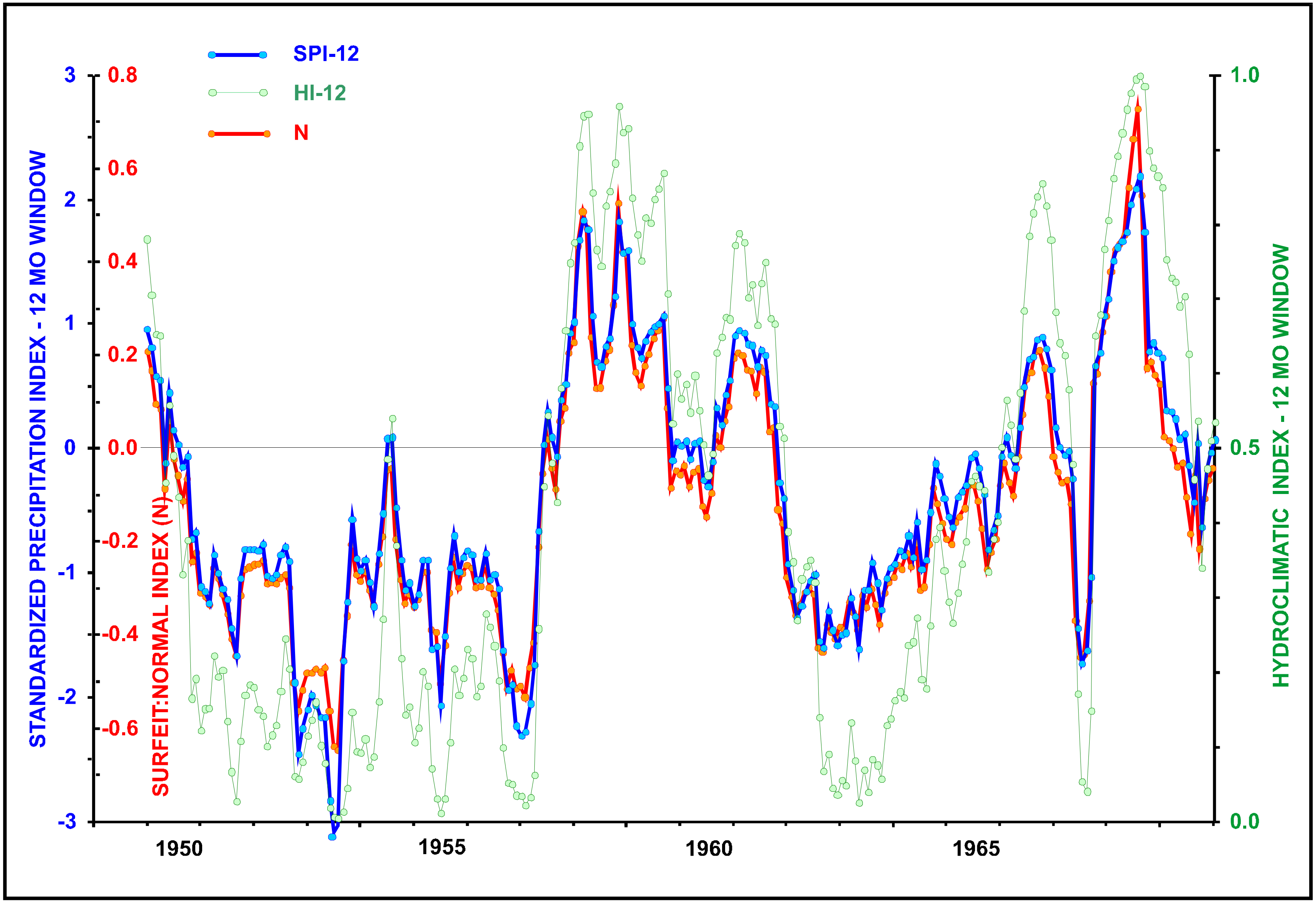

Figure 80 - Nueces River at Cotulla moderate (1 year) memory indices (see Table 17), for 1950-1969

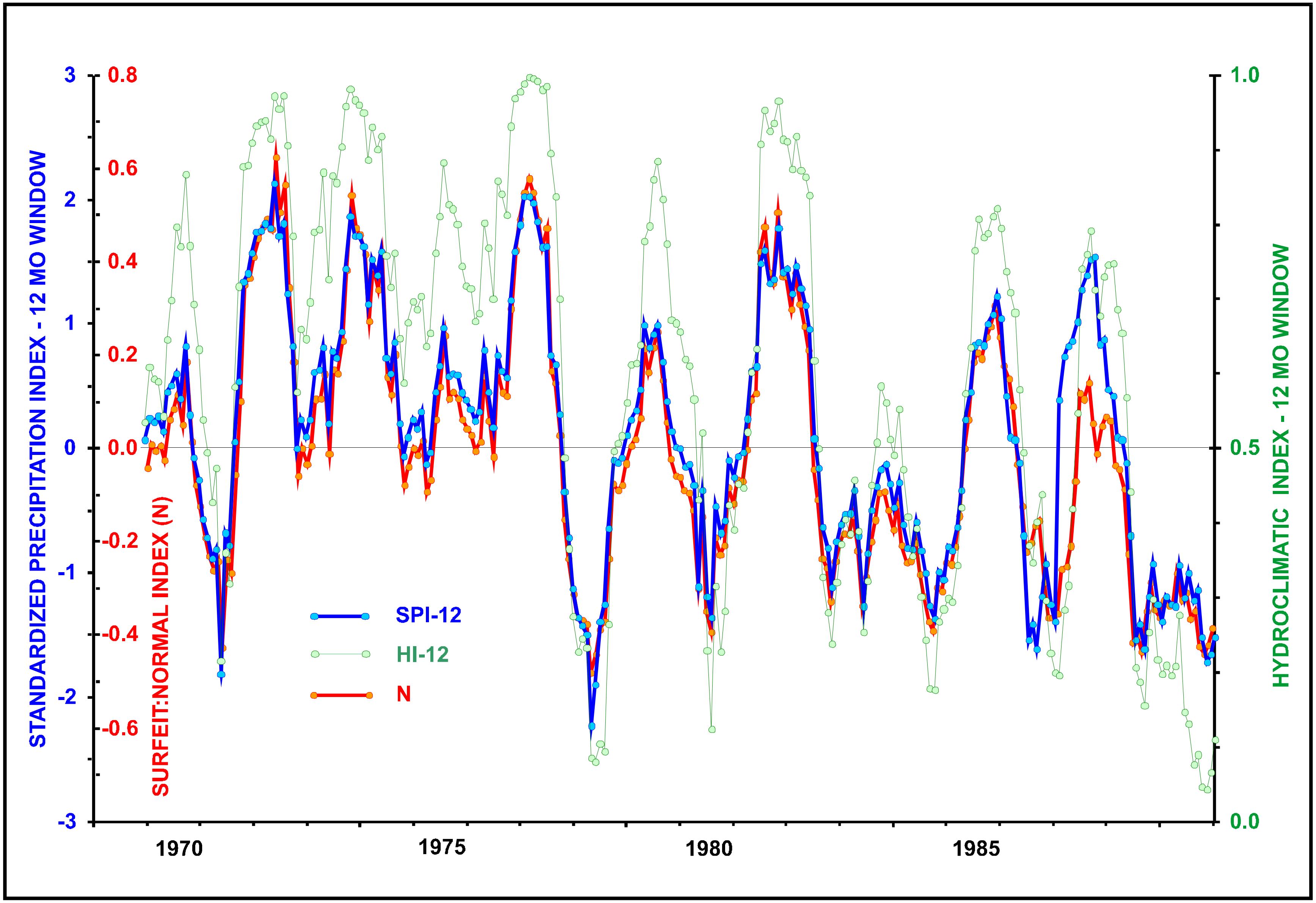

Figure 81 - Nueces River at Cotulla moderate (1 year) memory indices (see Table 17), for 1970-1989

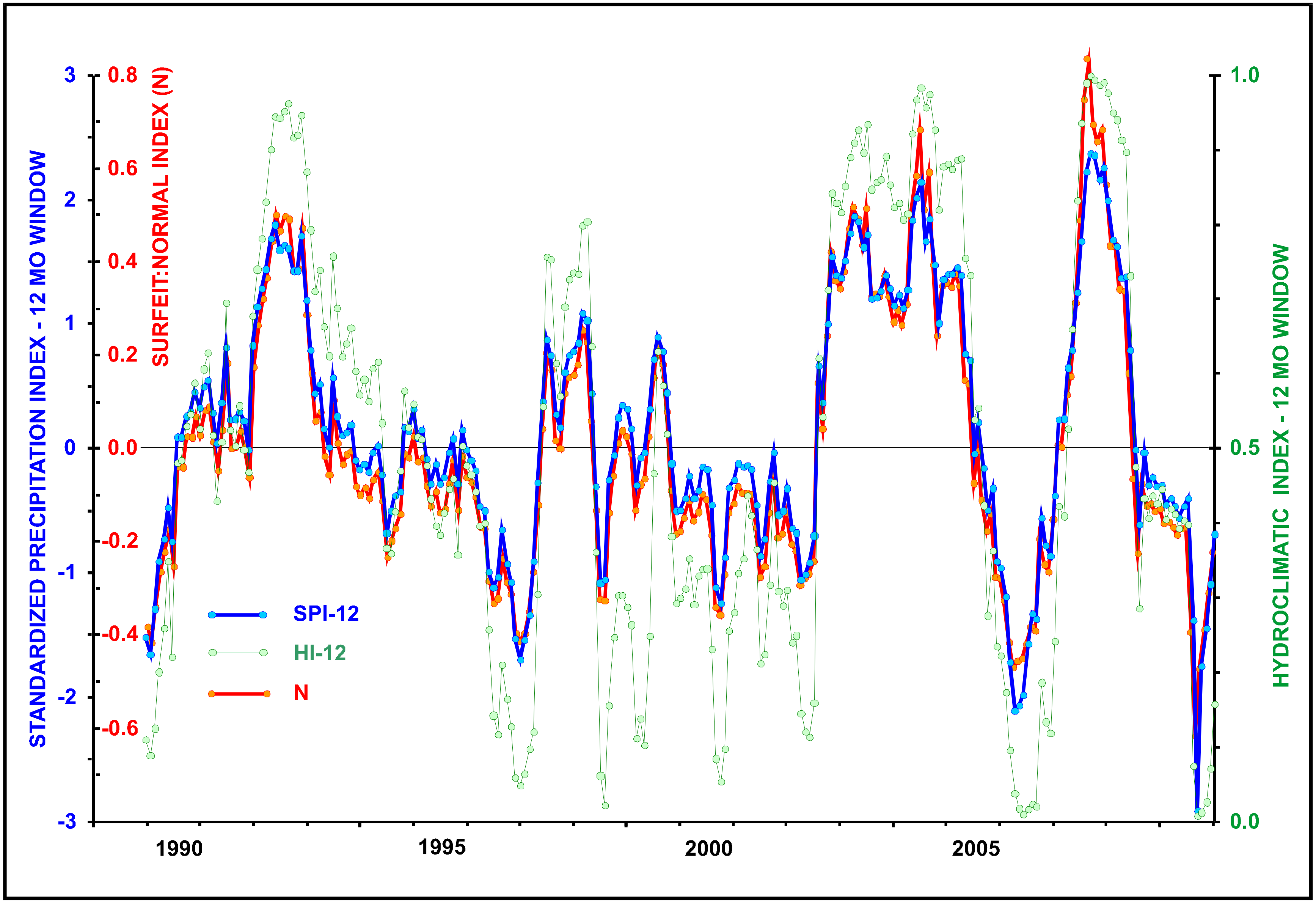

Figure 82 - Nueces River at Cotulla moderate (1 year) memory indices (see Table 17), for 1990-2009

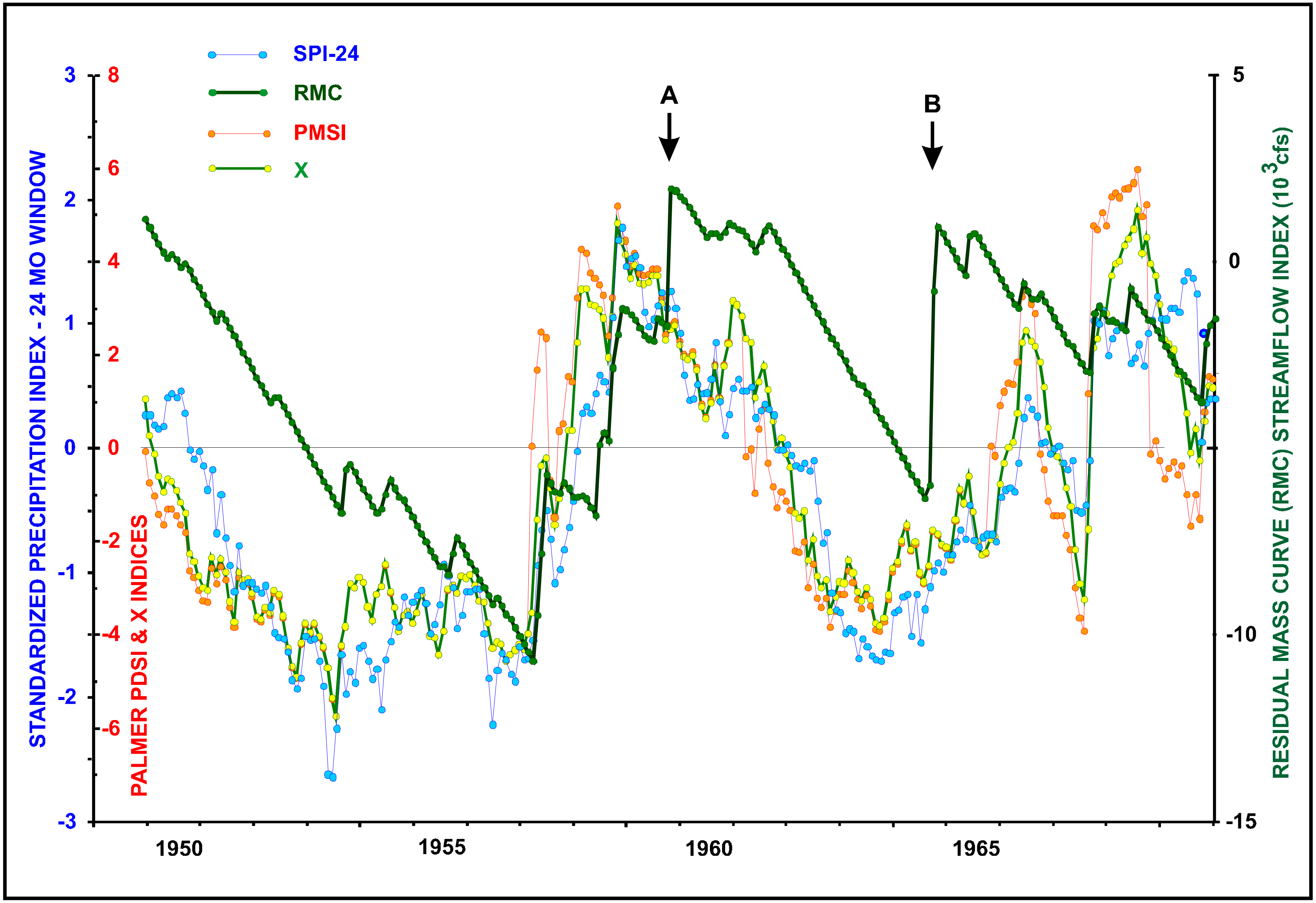

Figure 83 - Nueces River at Cotulla long-term memory indices (see Table 17), for 1950-1969

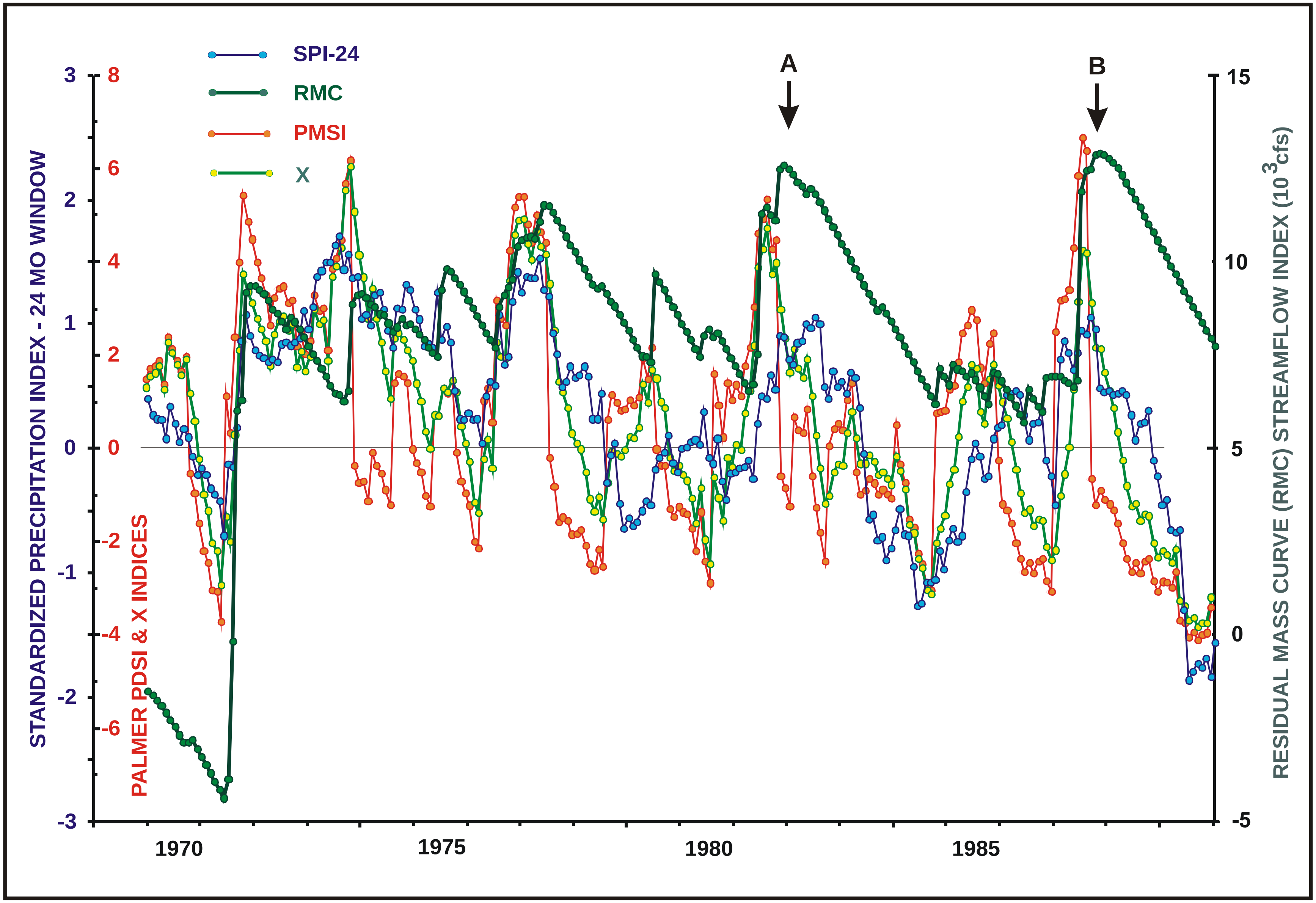

Figure 84 - Nueces River at Cotulla long-term memory indices (see Table 17), for 1970-1989

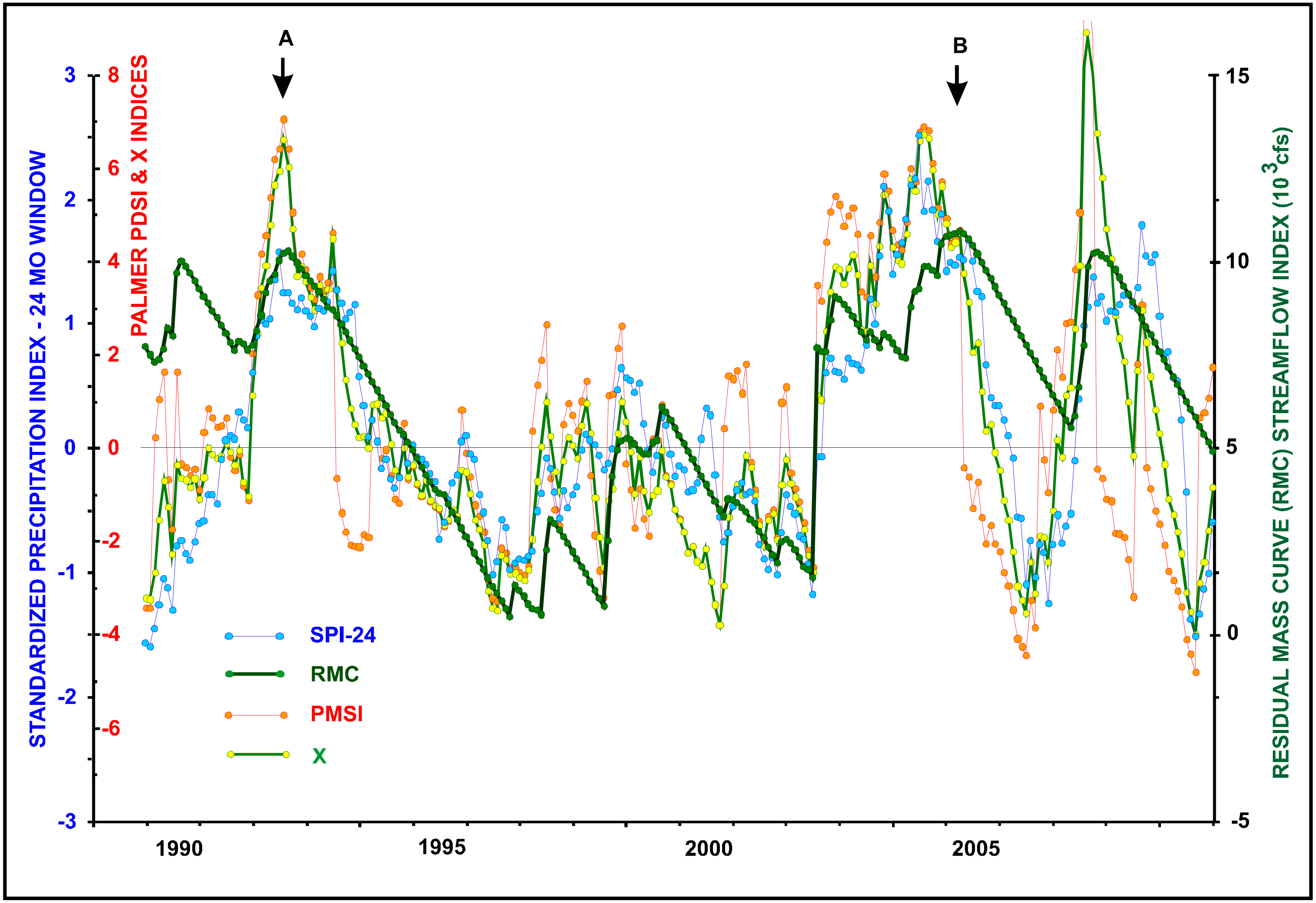

Figure 85 - Nueces River at Cotulla long-term memory indices (see Table 17), for 1990-2009

4.10 Observations

The short-term memory indices are not expected to be of particular interest in Texas water management, but are included in these test examples for completeness. HI-6 exhibits oscillations over its range of values with approximately annual period. This appears to be the result of HI-6 being sensitive to the annual cycle of *PE*, particularly when it is in phase with the semi-annual cycle of precipitation, that is, when the low-rainfall summer coincides with the high *PE* of summer. A randomly sampled example is shown in Figure 86, in which the negative of *PE* is plotted along with *P – PE*. This sensitivity is compounded by the sum over six months, which operates like a sliding 6-month mean. For this reason the excessive oscillation in the HI-6 renders it less effective an indicator compared to SPI-6 of short-term moisture variability. SPI-6, however, also exhibits excessive oscillation, being sensitive to the seasonality of rainfall.

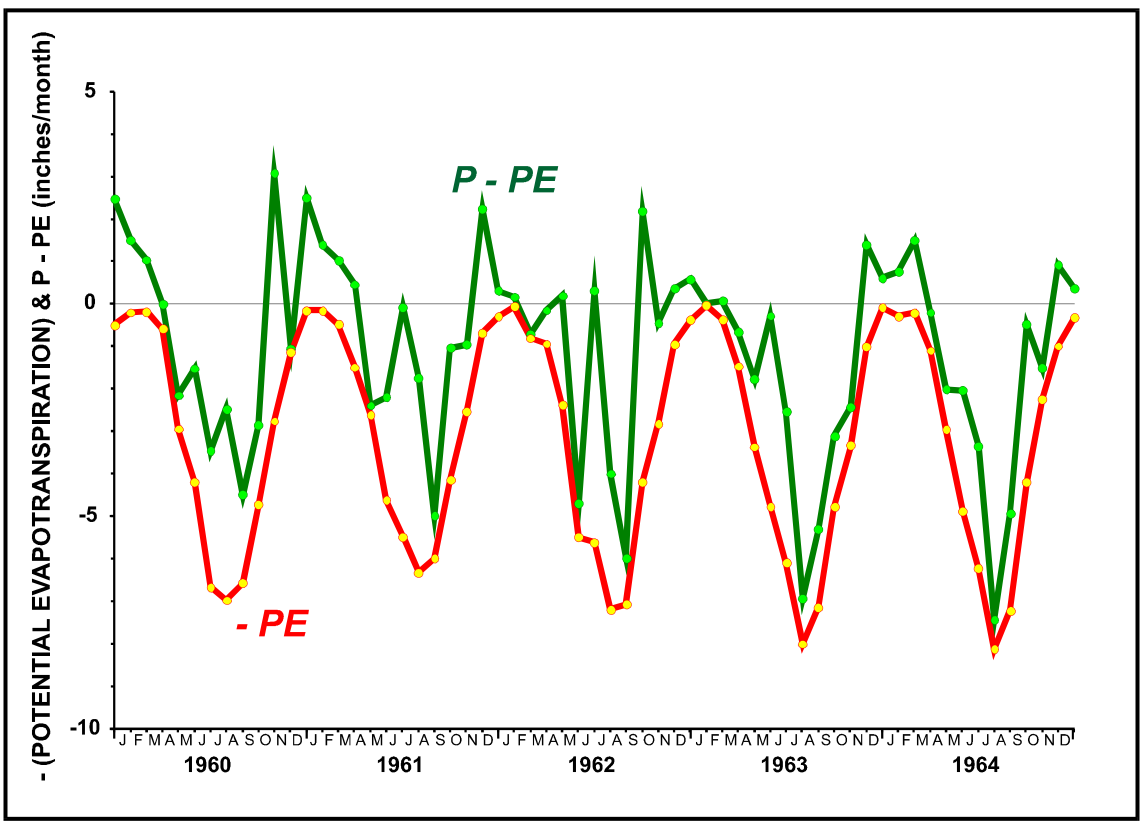

Figure 86 - Potential evapotranspiration (plotted as – *PE*) and *P – PE* for Brazos River at Seymour (NCDC Division 2)

Utility of the correlation coefficients (e.g., Table 18) in this evaluation is to exhibit the differences in moisture response of the various indices. Correlation among indices, either high or low, is not a concern in identifying potentially useful indices. It serves more as a measure of information content. Two indices with high correlation provide essentially the same information to the user, so the choice between them would be based upon other factors, such as ease of computation. Two indices with low correlation, on the other hand, provide different information. Whether one (or both) are useful must be decided by other measures. This poses a problem: there is no standard for moisture conditions against which a candidate index can be evaluated.

Three indices with 12-month memory were examined, *viz*. SPI-12, HI-12 and N, defined by (36). The most notable feature of these indices is their high correlation, see Table 18 *et seq*., all exceeding 0.90 and most exceeding 0.95. The simplest of these is the N index, but because SPI-12 is routinely provided by NCDC for each division in the state, use of SPI-12 is probably the easiest. SPI-12, it will be recalled, is based entirely upon precipitation, while HI-12 includes the water demand as measured by potential evapotranspiration (PE). (The strong seasonal oscillation noted above for HI-6 is averaged out when the 12-month index is used.) The correlation between these two indices, which is highest for the most humid regions (e.g., the Sulphur and the Little Cypress) and lowest for the most arid regions (e.g., the Brazos and the Nueces) indicates that including PE in the index does not markedly improve its explanatory power, because the two indices contain essentially the same information. This is suggestive that the SPEI-12 (Section 3.1.4.3) would not perform appreciably differently from SPI-12.

It is noteworthy that the highest correlation of all of the pairwise values is between N and SPI-12, ranging 0.97-0.98 over all the gauges and divisions tested, Table 18 *et seq*. This means that, apart from scaling, the two indices are essentially identical. Despite its apparent sophistication in applying a rigorous distribution transformation to convert the index values to a standardized Gaussian, the SPI-12 does not improve over a simple 12-month average related to the mean normal rainfall. This also suggests that both indices are equally capable of normalizing for geographic variation.

Despite the high correlation among these 1-year indices, there is occasional aberrant behavior, particularly of the SPI-12. Examples are “B” on Fig. 29 and “B” on Fig. 38. Of special note is the response of SPI-12 during the late 80's drought, in which it declares a non-drought, while the other indices continue to indicate drought, see "A" in Fig. 30, "B" in Fig. 39, and the same 1987-88 period on Figs. 51, 60 & 69.

To compare the long-term memory indices, it is useful to examine particular dry spells or wet spells, and to compare the corresponding 1-year indices, as well. The Drought of the Fifties is a prime example. This drought ran from 1950-51 to 1957-58, depending upon the region of Texas. The 1-year indices generally indicate a return to normal in 1953-54, so would consider this record drought to be made up of two component droughts, e.g. Figs. 29, 59, and 68. (On the Little Cypress and Brazos, Figs. 38 and 50, the 1-yr indices show two returns to normal, so this drought would be considered to be three shorter component droughts.) The long-term indices generally give a more realistic depiction of this event as a long, continuous drought of varying intensity, e.g. Figs. 32, 71, and 83. On the Brazos, Fig. 53, the Palmer indices correctly display the drought while the SPI-24 delays its onset by about two years. On the Little Cypress, Fig. 41, it is the PDSI that behaves erratically, recording two returns to normal during the drought, while the SPI-24 and Palmer X index behave more realistically. The RMC generally provides an unambiguous detection of the 1950's drought. In Figs. 44, 71, 74, and 83, there is a pronounced — and nearly monotonic — declining trend in the RMC. On the Little Cypress, in particular, where the PDSI, and to a lesser extent the X index and the SPI-24 are irregular, Fig. 41, the declining trend in cumulative flow from "A" to "B" clearly indicates a major drought. Only on the upper Brazos, Fig. 53, is the RMC ambiguous with respect to the drought, though it clearly identifies its start, agreeing with both Palmer indices and the 1-year indices (Fig. 50), but not with the SPI-24, which delays the onset about two years.

On these same figures, the four-year drought of the sixties, which is the drought of record for the Nueces, is clearly evidenced by the 1-year indices on the Little Cypress, San Bernard, Guadalupe and Nueces, Figs. 38, 59, 68 and 80, but much less so on the other test sites. The RMC indicates a drought starting around 1961 and ending around 1967 on the Sulphur, Little Cypress, San Bernard, Guadalupe, San Antonio, and Nueces, Figs. 32, 41, 62, 71, 74, 83, resp. On the Trinity, Fig. 44, the RMC indicates a seven-year drought beginning in 1958, "B", and continuing until 1966, "C", while the long-term indices show normal conditions until a short drought 1964-65.

On the upper Brazos, Fig. 53, the RMC indicates only a mild 4-5-year drought, in agreement with both Palmer indices, but not the SPI-24.

There are no prominent droughts indicated in the 1970-90 period, though there are relatively brief droughts in 1978-79, 1983-85, and 1988-89 confined mainly to the central and south-central regions of the state. In the 1970-75 period there is a pronounced pluvial in the Little Cypress, Figs. 39 and 42, San Bernard, Figs. 60 and 63, Brazos, Figs. 51 and 54 and Nueces, Figs. 81 and 84, for which the various indices, including the RMC, generally agree. The pluvial in the Guadalupe and San Antonio, Figs. 72 and 75, is more protracted, continuing until 1980, and is primarily evidenced in the RMC but not the rainfall-based indices.

In the 1990-2009 period, the most prominent moisture events are the intense droughts of 2005-06 and 2008-09, and the intense pluvial of late 2007. These are clearly displayed on the Sulphur and Little Cypress by all of the 1-year memory indices and by the long-term indices, Figs. 31, 34, 40 and 43. The RMC for these gauges, however, displays a protracted drought 2001-07, Figs. 34 and 43. The Guadalupe, San Antonio, and Nueces gauges show general agreement between the RMC and the rainfall indices.

In summary, while the SPI indices are conveniently updated and disseminated by NCDC, there may be advantage in exploring the use of the N index. Among the long-memory indices, the PMSI displays erratic behavior, with large excursions at the beginning or end of a dry spell. This behavior is not manifested by the X index, however, whose behavior to this reviewer seem more “realistic.” The RMC is the only index tested that is entirely streamflow-based. For a given streamflow gauge it would appear to offer promise as a moisture indicator for the watershed at that point in the stream channel. However, it is clear from a study of the demonstration cases above that this index is not yet in a form suitable for routine application.

5. DISCUSSION AND RECOMMENDATIONS

From the literature review documented in Chapter 3, seven indices with memories of one year or more (see Section 1.3) were selected for more detailed evaluation using data from Texas rivers, summarized in Table 17. Four of these are well-established in the literature, commonly used for hydroclimatological work, and appropriate for Texas. One, the X index, is easily derivable from the suite of Palmer indices routinely provided by the NCDC for the climatic divisions of the state. Two were devised in this study as variants on other indices — the N-index, which combines properties of the surfeit precipitation as a fraction of the mean and the SPI, and the RMC, which is a cumulative of streamflow — and therefore have little literature to support their use. Demonstrations of the performance of these seven indices are presented in Chapter 4 for eight river stations in the state representing a range of hydroclimatologies.

RECOMMENDATION 1: It is recommended that the state give consideration to using six of these, namely the PDSI, SPI-12, SPI-24, N index, X index, and RMC.

The SPI-12, the SPI-24, and the PDSI are recommended because they are widely accepted and these indices are readily available from, and routinely updated by NCDC. The X index is simply the PDSI stripped of its protocols for re-initializing based upon the start and termination of wet and dry spells. This index is readily computed from the Z index, routinely provided by NCDC, and equation (25). The N index is easily calculated from the time series of monthly precipitation. The RMC is likewise readily computed from monthly streamflow data, routinely available from USGS, and TWDB has already authored an EXCEL® application with VBA Macro to compute this index.

The SPI was one of the indices recommended by Quiring et al. (2007) for use in monitoring drought in Texas, specifically the SPI-12. (We also recommend SPI-24, which Quiring et al. did not.) The PDSI was also considered by Quiring et al. but not recommended because it "performed poorly." This was based mainly on an evaluation similar to that of Keyentash and Dracup (2002), in which six subjective criteria are assigned subjective scores and the composite score computed by subjectively weighting the six criteria. It is our judgment, also subjective, that the PDSI should be given consideration by the state. It has a physical basis that includes a rudimentary soil-water budget and may therefore yield information about pluvial or drought conditions that indices based solely on precipitation cannot, as suggested by the example of Figure 87. It has been widely used, especially during the intense droughts of the past decade, and both the public and water management professionals have acquired a level of comfort with the index. In the preparation of the Senate Bill 3 BBEST recommendations for the Colorado and Gaudalupe basins, use of the PDSI as the criterion for determining whether wet, normal or dry conditions prevailed was the consensus proposal. The PDSI was used in a recent report to TWDB to correlate with salinity in determining drought impacts on Texas estuaries (Montagna and Palmer, 2012). This is not to say that the PDSI is without problems. These are summarized in Section 3.1.3. In particular, the protocols for modifying the index when dry spells or wet spells begin and end produce erratic responses in the PDSI. These are avoided by using the X-index, hence its recommendation to the state for consideration.

The HI-12 index was intended to represent the suite of precipitation – evapotranspiration indices (Section 3.1.4), of which the SPEI is presently the most prominent but is too complex to consider for use in the present application, see desideratum (vi) of Section 1.1.1. HI-12 proved, however, to be redundant, being almost perfectly correlated with the SPI-12 and N indices. It would appear little improvement in the index results from the additional complexity of including PE in its calculation. Similarly, the SP-12 and N index are highly correlated. But the choice is not as simple. They are highly, but not exactly, correlated, and occasionally the SPI-12 appears erratic compared to N. An example is its 1987 response to in increase in precipitation, see "A" in Fig. 30, "B" in Fig. 39, "A" in Fig. 51, etc. On the San Bernard, Fig. 60, this SPI-12 "event" ends a drought that the N index indicates lasting until 1989. More study is required to determine the reasons for the differences in the SPI-12 and the N index, and which one is preferable.

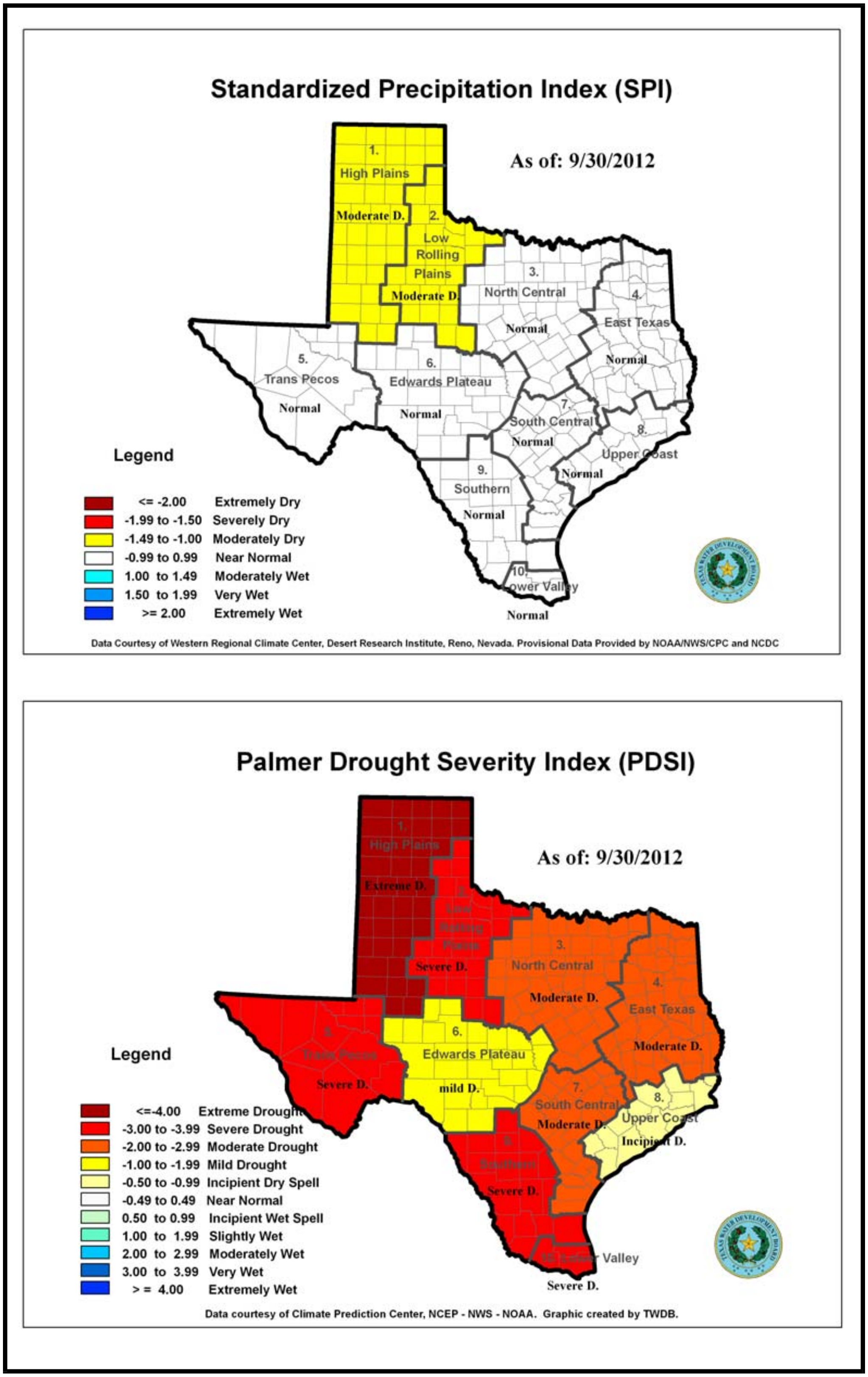

Figure 87 – Drought conditions in Texas indicated by the SPI (above) and PDSI (below), from Drought Preparedness Council (2012)

RECOMMENDATION 2: It is recommended that the state further develop the RMC, to identify and explicate its behavior, devise a scaling methodology to normalize the index to a more useful range, inquire into a decaying-memory formulation, and explore alternative methods for determining the start and termination of drought and pluvial periods.

The RMC is the only streamflow-based index recommended. It is simple, and appears to be capable of identifying long-term droughts that are missed by the climatological indices discussed above. It is also noteworthy that it exhibits low correlation with the other index variables, see Table 18 *et seq*., which suggests that it provides a different class of information about streamflow behavior than can be inferred from the meteorological indices. Its formulation includes intuitive measures of intensity of drought and pluvials, see Fig. 25 and Table 14. (As noted earlier, this study was underway during the Guadalupe BBEST effort, and this index was suggested as a method for identifying historic droughts in the inflows to San Antonio Bay. The TWDB EXCEL workbook implementing the RMC using Macros was recently employed by Montagna and Palmer, 2012, to determine droughts in inflows to the principal Texas estuaries.) There are, however, apparent inconsistencies in its behavior from one gauge to another and at different periods of time for the same gauge. Certainly, part of this is the totality of upstream influences on streamflow, in addition to hydroclimatology. The fact that in its present form, the index is not scaled, but is a simple cumulative of the preceding data record, leads to a wide range of values, see the demonstration cases in Chapter 4.

Because the residual mass curve of streamflow is a cumulative, the kernel $h_i = 1$ for $i = 1, 2, \ldots, N$, i.e., extending over the period of record. Thus, all historical flows have the same influence on the index. Though this index has great diagnostic value (see the introductory paragraphs in Chapter 3), some workers find this property undesirable. A kernel can be specified that decays with time, thereby shortening the memory of the index. Though no examples of this could be found in the literature, it warrants research.

RECOMMENDATION 3: It is recommended that the state compile various data quantifying drought and pluvial impacts and use to "calibrate" the various indices to numerical categories based upon actual drought or pluvial conditions. This will remedy a major weakness in the utility of all of the above indices, and any additional indices that might be entertained in the future.

Though the use of these indices can be implemented at once — and in fact several of these are already in use by TWDB as part of its support of state drought management activities — none of these indices is immediately able to serve as a *comprehensive* metric for the state in water monitoring or water planning without the additional research and development recommended above. This is because the available moisture-condition threshold categories, tabulated where available throughout Chapter 3, are arbitrary and provide only a relative indication of hydrometeorology. A truly comprehensive metric would connect moisture conditions quantitatively to moisture surfeit or deficit impacts in the state. Because this motivates the above recommendation for direct research, some background is necessary.

There are five steps, or components, in the development and practical application of a moisture index:

(A) formulation of the index, i.e., specification of its functional dependency on hydro-climatological variables;
(B) scaling of index values to a useful range of variation;
(C) identification of scaled values with moisture surfeit or deficit conditions;
(D) definition of a protocol for identifying moisture surfeit or deficit *events*; and
(E) specification of thresholds for different moisture categories.

Component (E) can then become the basis for establishing triggers to which specific management actions are tied. From the literature review of Chapter 3, it is apparent that for most of the indices examined, not all of these steps have been carried out. This results in disconnects between the computation of an index and its practical application in assessing moisture conditions. In the literature, the emphasis for most of these indices has been on components (A) and (B). This fact is responsible for one of the most disappointing results of the literature survey, that no cases could be located in the international literature reviewed in this study in which a set of definitive, physically-based triggers is used in practical management of water resources.

As an example, consider the SPI, described in Section 3.1.2. The basic idea is to form an index that is "standardized" by dividing a moisture anomaly, i.e., the amount that a measure of moisture is above or below its normal, by the standard deviation of the measure, analogous to how one transforms a Gaussian-distributed variable to the standardized Gaussian with zero mean and unit standard deviation. The measure in this case is a sum over M months of monthly rainfall, and to account for skew in these sums, they are fitted to some standard skewed distribution, which is then transformed mathematically to a standardized Gaussian. In other words — and cutting through the jargon — the SPI-M takes a record of monthly precipitation, creates a new record of M-month sums, and produces an index that generally ranges $\pm$ 3 about 0 (though theoretically its range is unlimited). Thus steps (A) and (B), above, are completed.

Step (C) for the SPI, however, is problematic. Moisture conditions for the SPI are usually demarcated in the literature by multiples or half multiples of the (unit) standard deviation, given in Table 7, a step which combines component (C) and (E) by, in effect, using (E) as calibration for (C). There is no physical basis for the moisture categories (other than the vague notions that half of the time conditions are anomalously dry and the other half they are wet, and that "severely dry" is less common than "moderately dry" and therefore should have a more negative index, etc.). The categories are merely convenient integer (or half-integer) values of the index. While most users of the SPI simply adopt these recommended categories, some users have attempted more resolution in the moisture categories, such as those of Table 16. These, however, still lack a physical basis for the assignment of categories. This lack is manifested in both the descriptions of the categories ("moderately wet," "severely dry") and the arbitrary assignment of integer index values. Indeed, one may ask why a precise demarcation of the index should be needed if the moisture category is so vaguely described.

Another example is the so-called precipitation-decile index of Gibbs and Maher (1967) described in Section 3.1.1. This employs a simple indicator, namely annual or monthly rainfall, whose cumulative frequency *is* the index, with vague categories of moisture condition assigned to whole multiples of 10%, see Table 5. Again, qualitative, subjective categories are used to "calibrate" the association of the index of cumulative frequency with moisture conditions. Only to this limited extent is component (E) accomplished. We note that, despite its simplicity and lack of a physical basis for the threshold categories, this index has had considerable success in communicating dry or wet conditions to the public in Australia (e.g., Kininmonth et al., 2000).

The precipitation-decile index illustrates two important points about indices in general. First, this type of index is not unique to precipitation. Almost all of the indices considered in this review can be easily converted to a cumulative frequency depiction (the exception perhaps being the RMC for streamflow), and then moisture categories assigned to specific cumulative frequencies. Any index for which this can be done is therefore a "quantile index" of which the decile index is a special case. This effectively replaces the defined magnitude of the index with a cumulative frequency, i.e., it re-scales the index to [0, 1] (or, if one prefers percents, to [0, 100]). An example for Texas is shown in Figure 88, showing ogives for indices tested in Chapter 4, plus the "percent-of-normal" index (Section 3.1.1). Steinemann (2003) recommended exactly this tactic of transforming an index to cumulative frequencies to remedy the confusion of different index ranges, and to better compare the consistency between indices. However, this does not remedy the problem of connecting categories to physical manifestations of moisture condition. Defining categories by even multiples of 10% (or 5%, or 20%) has no more significance than even multiples (or half multiples) of a standard deviation, say, unless there is a separate defensible physical basis.

Second, ordering the values of the index (monthly precipitation, say, or their *M*-month sums) by magnitude to display their cumulative frequencies is a useful exercise to communicate the relative wetness or dryness of the individual data points, but this ignores the time sequence of these data points. This motivates the next recommendation, and is further discussed in that context.

We have seen that most indices in fact prove to be arbitrary in the association of the index value with physical moisture conditions, mainly because assignments of thresholds of wet or dry conditions are themselves arbitrary. (The Palmer index is an exception. Palmer, 1965,

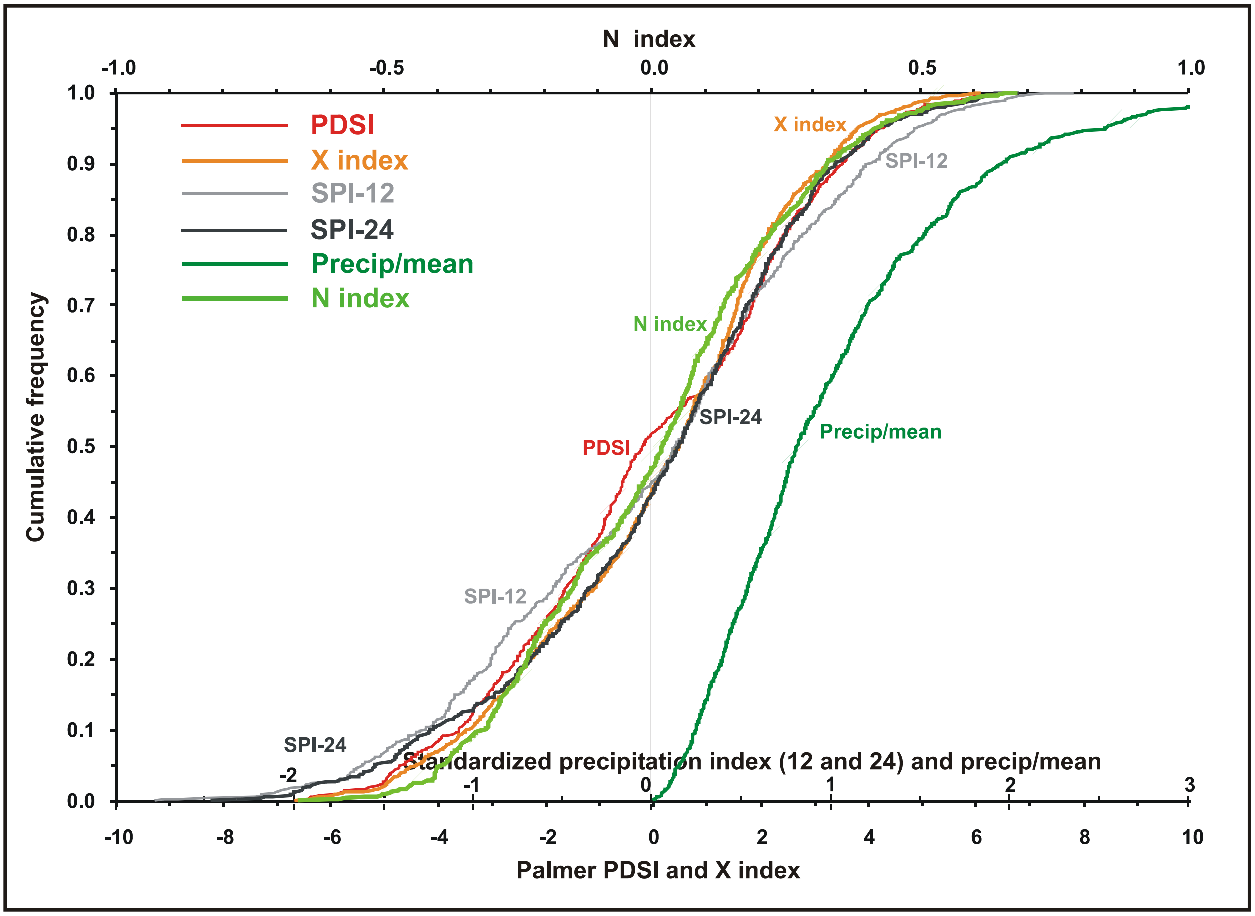

Figure 88 - Empirical cumulative frequency distributions for selected indices, NCDC Division 7 (South Central Texas), 1941-2011

developed his indices from a study of drought conditions in the Midwest, scaling these to severe drought conditions, then interpolating the PDSI values for intermediate conditions.) This deficiency in a moisture index, particularly for drought, has received some limited research attention in recent years.

The Drought Monitor project (Svoboda et al., 2002) has proposed a set of thresholds of drought impact and associated cumulative frequencies, without an associated index. These are given in Table 26. Only drought conditions are defined, but the scale could be extended into the water-surfeit categories. The Drought Monitor team has established correspondences between these

Table 26
Drought magnitude categories of Drought Monitor, and associated impacts see Svoboda et al. (2002)

category	*condition*	*probability**	*agriculture*	*water supply*	*fire risk*
D0	abnormally dry	(0.20, 0.30]	slowed activity & crop growth	streamflow below average	above average
D1	drought—moderate	(0.10, 0.20]	some damage to crops & pasture	flow, lake & well levels low, some shortages	high
D2	drought—severe	(0.05, 0.10]	crop & pasture losses likely	shortages common, restrictions imposed	very high
D3	drought—extreme	(0.02, 0.05]	major crop & pasture losses	widespread shortages & restrictions	extreme
D4	drought—exceptional	≤ 0.02	exceptional & widespread crop & pasture losses	shortages create emergencies	exceptionally dangerous

* These define the relative rarity of conditions for a preset increment of time, so that the probability is interpreted as the chance of those conditions occurring in 100 such time increments. For annual events, for example, this would be the frequency of occurrence per 100 years.

categories and drought classifications for impacts on agriculture, water supply and wildfire, also shown in Table 26. Unfortunately, without a quantification of these impacts, it does not provide the level of definition that is necessary to effect a direct translation between the monitor category, or, for that matter, the value of any of the indices, and the associated impacts. What, for example, is the distinction between "common" and "widespread", or "extreme" and "exceptionally dangerous"? But this does provide a framework so that as specific data on impacts are collected, the impact categories should become more objective and quantitative. For instance, in order to develop meaningful thresholds for the Arizona Governor's Drought Task Force, Goodrich and Ellis (2006) adopted the categories of the Drought Monitor (Table 26), determined thresholds of the PDSI and SPI that correspond to the Drought Monitor frequencies, and analyzed three historic drought periods in terms of these indices with the severity categories of the Drought Monitor.

There are a handful of recent research results that seek empirical relations between agricultural data and moisture indices. Agriculture is particularly attractive as an indicator of drought impacts because of the extent and reliability of numerical data. Chipanshi et al. (2006) used the rainfall-decile index threshold method in which the decile categories were roughly validated by crop condition and vegetation vigor reports on the Canadian prairies. Further studies are underway in Canada to better assess the correspondence between the index and actual drought impacts. For example, Sun et al. (2011) used data on spring wheat yields, de-trended and standardized, to quantify impacts of drought on agriculture. These were then used to evaluate the wheat indices corresponding to the cumulative frequency thresholds of the Drought Monitor (Table 26). In Greece, Mavromatis (2007) compiled data on wheat harvests, and evaluated their correlation with several versions of the Palmer index and with the SPI. (Mavromatis does not state the aggregation time for the SPI, but did apply a 12-month sliding average to all data sets before determining their correlations. The PDSI was found to provide the best overall performance.) Patel et al. (2009) related the SPI to food-grain production anomaly in India.

Certainly agricultural data should be part of the data compilation of Recommendation 3, because this type of data is acquired routinely in Texas and is available historically. But this data also suffers a severe disability, in that the data does not respond only to moisture conditions but also the strategies of farmers and ranchers in reacting to economic forces as well as climate. Herd size or winter planting will be different for a given moisture state in the second year of drought versus the fifth year. Similarly, during pluvials the mix of crops will vary over time depending upon supply and demand. Some measures of moisture condition impact uncorrupted by extraneous considerations are therefore needed. In this context, several promising technologies, which have been ignored in this review because they fail desideratum (vii) in Section 1.1.1 of an extant historical data base, may be of considerable value. These include soil moisture measurements and remote sensing of vegetation, such as the vegetation condition index derived from AVHRR imagery. Data on soil moisture, together with precipitation, temperature, PDSI and related meteorological variables, and on crop production are gathered by the Joint Agricultural Weather Facility, an operational unit under the auspices of the U.S. Department of Agriculture and the Climate Prediction Center of NWS (Motha and Stefanski, 2006). A project underway at TAMU to compile a North American Soil Moisture Database (see soilmoisture.

tamu.edu) would clearly be an essential resource. Sullivan and Maidment (2012) recently compiled a GIS-based mapping of soil moisture in Texas drawn from the North American Land Data Assimilation System (NLDAS) re-analysis project, which might serve as a suitable proxy data source. It is also possible that model-based methods such as those promoted by Sheffield et al. (2004), Andreadis et al. (2005), Narasimhan and Srinivasan (2005), and Meng and Quiring (2008) might enable a synthetic historical data base to be created where field data are wanting. All of these possibilities would of course need to be evaluated in the suggested research.

RECOMMENDATION 4: It is recommended that the data compiled as a result of Recommendation 3 be applied to re-evaluating the protocols for determining the start and end of anomalous moisture periods.

It was observed above that, while the construction of an ogive by re-ordering indicator values by magnitude has utility, it overlooks a fundamental aspect of the hydroclimate, namely the time sequence of that indicator. That is, the cumulative frequencies provide no insight into the occurrence of spells of dry or wet conditions. It is easy to construct examples of an index with the same distribution of magnitudes but different time sequencing to produce different pluvial or drought occurrences. In addition to the definition of the index itself, some protocol must be provided by which that index is used to determine the beginning and end of a spell of anomalous moisture conditions. This is component (D) above. Many of the indices reviewed in Chapter 3 do not have such a protocol, and their practical utility is diminished as a consequence. Notable exceptions are the PDSI, the SPI and the Australian precipitation-deciles index. However, users of these indices sometimes ignore the associated drought protocol, and simply compare index values to other measures, to arrive at a judgment on how well (or poorly) the index represents drought conditions, e.g., Naresh Kumar et al. (2009) for the SPI, and Dahm et al. (2003) for PDSI. (It can be argued that since PDSI includes drought and pluvial detection as part of its computation, the categories should be sufficient to identify anomalous spells.)

The fact is that the protocols for delineating anomalous wet or dry periods are as arbitrary as the moisture-condition categories. Once the data base of Recommendation 3 is available, this data can also be applied to devising better, physically-based protocols for indentifying drought or pluvial periods in the time series. This would also permit the study of additional aspects of anomalous moisture conditions, *viz*. seasonality, periodicity, persistence, and spatial distribution. The review of Beran and Rodier (1985) is useful for its breadth delving more deeply into these and related topics with examples drawn from the climatological literature.

RECOMMENDATION 5: It is recommended that if any indices are selected for use by the state that involve equiprobability transformations to a standardized Gaussian, the specific standard skewed distribution to be employed be given detailed study.

For those indices like the SPI (see Section 3.1.2.1), the selection and use of non-Gaussian distributions to fit hydrometeorological variables need to be rigorously addressed. This is central to the SPI and some of the other indices that use a similar aggregation strategy, such as the SFI and SPEI (which are not recommended for use by this study), whose standardization method involves the inverse Gaussian transform of the fitted non-Gaussian distribution. It has been amply demonstrated in the literature reviews of Chapter 3 that there are many skewed distributions that will serve this function, and that the specific values of the index varies with the selected distribution. It has also been demonstrated that there is little statistical basis for selection. More rigor is needed in the fitting procedures, including statistical assessments of confidence bounds on the fitted curves. (In this context, Fig. 88 should be re-examined. These empirical ogives were plotted on axes specified only to center the graph and to limit or expand the range for readability. The clustering of ogives is trying to tell us something.)

In addition, if time scales on the order of 1 to 6 months are used for the SPI, as recommended by Quiring et al. (2007), then, as described in Section 3.1.2.2, an issue arises if there are months with zero values of the variate S (cumulative streamflow or precipitation). This issue becomes increasingly likely with distance south and west in Texas. In the present procedure for the SPI, these data are excluded from the analysis for fitting the gamma function, then restored before the resulting ogive is inverse transformed to the standardized Gaussian, see equation (11). For application in Texas, research is needed to determine how serious and widespread this problem might be and, if necessary, devise an alternate strategy, either a fit to the gamma function that is nonsingular at zero, or selection of a different distribution.

It will be recalled that one of the purposes of such standardization of an index is to render it independent of geography (see Section 3.3.1). Much has been made in the literature of the need for an index that normalizes variation in space, so that a wet or dry condition becomes a matter of scaling the index to local conditions. But the utility of normalization in space must be given careful philosophical consideration in the context of the intended use of the index. In their critique of the SPI, Lloyd-Hughes and Saunders (2002) observe that its standardization means that the threshold values for wet or dry conditions will occur with the same frequency at all locations (over a long period of record). This means that the index will not be able to identify "drought-prone" or "fluvial-prone" regions. This is, in fact, a weakness of any standardized index, to the extent that that standardization is successful, including the PDSI. Clearly, if identification of such regions is a goal, then the strategy of normalization must be carefully re-considered.

RECOMMENDATION 6: It is recommended that the use of reservoir contents both as a statewide or regional indicator, and as an index for management of water demand be given more extensive and rigorous study.

Reservoir contents may be a poor indicator of overall moisture conditions, because it responds specifically to conditions on its watershed, may be affected by upstream storage and diversion, and may be depleted due to factors unrelated to hydroclimatology. However, as mentioned in Section 3.3, there is much to recommend reservoir contents as a practical indicator for water-demand management. The lack of cases on physical connection of water demands to hydroclimatological indices is a disappointment of the literature review. Nonetheless, this is a topic that should be pursued with directed research. It is also suggested that the TWDB augment its present statewide reservoir contents accounting by providing separate percent-of-capacity for water supply reservoirs and for power reservoirs. (Multipurpose reservoirs with both functions should be included in both categories.)

RECOMMENDATION 7: It is recommended that appropriate thresholds for applicable indices be developed to differentiate the broad categories of "wet," "normal," and "dry" used in the Senate Bill 3 environmental flows recommendations.

While exploration of potential indices for use in categorizing streamflow conditions for environmental-flows purposes was one of the motivations for the present study, its scope did not extend to apply a candidate index for this purpose. Indeed, the development work of Recommendations 3 and 4 as applied to aquatic ecosystems would need to be completed before specific application to environmental flows specification should be undertaken. It would be desirable, however, that such development proceed so as to dovetail with the early projects in the BBASC work plans, so that this information can be exploited as a part of adaptive management of these flows.

RECOMMENDATION 8: It is recommended that techniques for drought prediction of potential applicability to Texas be given exploratory study based on literature review and experimental testing in selected regions of the state.

Nalbantis and Tsakiris (2009) list several desiderata for hydrometeorological indices. These were addressed in Section 1.1.1, except one, namely:

(xi) capable of revealing potential important moisture conditions within a short time lag after their occurrence.

(to continue the numbering from Section 1.1.1). The desirability of such a property is clear. It could provide early warning of the start of a severe drought, or signal the end of a prolonged dry spell. But this also invokes an aspect of an index that moves into an entirely new dimension, namely a tool for prognostication. The Texas Water Development Board expressed an interest in the utility of an index in forecasting river flow over a forthcoming season (see technical objective 4 in the Introduction). This is in effect desideratum (xi) above. This proved to lie beyond the resources of the present study, for reasons that will be made apparent, but remains a potentially important topic deserving additional targeted research.

Fundamentally, the forecasting of moisture conditions in general, and river flow in particular, requires first a forecast of future weather. One of the seminal results in atmospheric dynamics of the latter part of the twentieth century is the discovery that the dynamic methods employed in meteorology for forecasting weather are essentially chaotic, so that there is a limit to the predictability of the atmosphere. This limit is now considered to be one to two weeks, depending upon location and the state of the atmosphere. If predictions for longer time frames are possible, they cannot be dynamical, but must entail statistics. There are two basic strategies: (1) extrapolate hydroclimate from past conditions to a future horizon, (2) relate hydroclimate to larger-scale modes of the atmosphere, and attempt to forecast these modes to a future horizon. Examples of the former range from simple regression of the forecast variable on time, to Markov models, and autoregressive integrated moving average models (e.g., Steinemann, 2003; Cancelliere et al., 2007a; Nalbantis, 2008). Examples of the latter include seeking the relation between rainfall and streamflow on the El Nino-Southern Oscillation (ENSO), North Atlantic Oscillation (NAO) and Pacific Decadal Oscillation (PDO), among others (e.g., Muñoz-Díaz and Rodrigo, 2005). Within each strategy, there is a myriad of techniques reported in the literature. Some of these have employed hydroclimate indices as a convenient quantitative summary of the complex of atmospheric and hydrologic processes.

A good example of the use of statistics to establish probabilities of drought conditions changing is the study of Steinemann (2003) in the Apalachicola-Chattahoochee-Flint river basin, the focus of the water-allocation dispute among Alabama, Georgia and Florida. Both the SPI and the PDSI were modeled as a six-state Markov process, the six states corresponding to the drought triggers used in the basin to characterize moisture conditions. The continuation or intensification of an ongoing drought was represented by the transition probabilities in any given month that the next month would be in the same category or worse. Therefore, in principle, the probability of a major drought continuing was dependent upon the historical behavior of the index. Another example is the drought severity assessment for Greece of Vangelis et al. (2011). This was based on the RDI (Section 3.1.4.2) in which both the numerator *P* and denominator *PE* were assumed to be Gaussian distributed random variables, so that the RDI could be depicted as a joint bivariate distribution, from which a return frequency can be estimated.

There is a prodigious literature on coupled modes of variation of the atmosphere and ocean, of which the IPCC report provides an excellent summary and guide to the literature (see IPCC, 2007, especially Section 3.6). Despite large-scale association between ENSO and NAO and regional rainfall, these indices are not reliable forecasters of moisture conditions or streamflow in Texas (e.g., Ward, 2010; Slade and Chow, 2011). However, they may prove useful with more sophisticated statistical studies, such as pursued by Cordery and McCall (2000) in Australia.

Perhaps the most promising candidate for forecasting in Texas is an example of a statistical technique that exploits seasonality of rainfall and streamflow. In the Mediterranean, this has been employed in Cyprus (see Section 3.3.2.1), Spain, and Greece for limited seasonal forecasting of drought conditions. The extent to which wet-season rainfall (winter-spring in the Mediterranean) is reduced below some norm is used to forecast water shortages in the subsequent low-flow season (summer, in this case). Many regions of Texas exhibit seasonality in rainfall, see Fig. 2, that might be similarly employed.

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APPENDIX

Review comments of the Texas Water Development Board: revisions, reconciliations and responses

TWDB — General observations:

In this draft final report, several descriptions of flow conditions ("indicators" or "indices") that may be useful in Texas are identified. Descriptions of the indicators and indices are incomplete. Test sites for evaluating and demonstrating techniques are identified in the report, but results and conclusions are not included. A significant amount of additional documentation is required in order to transform the draft report into a suitable final report. The draft report does provide a suitable outline for an adequate final report.

Response:

Regrettably, the product reviewed was an early draft. The observations above are exactly correct. The revised report is an order of magnitude larger, therefore contains a significant amount of additional documentation, and remedies (hopefully) all of the above deficiencies.

TWDB — Required change number 1:

In order to be factually correct, the first sentence of the first paragraph on the first page needs some modification. The phrase "within the past decade this concern acquired a legislative focus" should be modified to "within the past twelve years this concern acquired a legislative focus."

Revisions:

This first section was overly long and dwelled too much on environmental flows, in comparison to the other motivations for the study, so in revision it has been substantially reduced. The offending sentence has been excised.

TWDB — Required change number 2:

2nd sentence, 1st paragraph, 1st page. The phrase "Seasonal flow variation considered necessary" should be modified to "flow variation considered necessary." TIFP is interested in inter-annual, as well as seasonal variation. The interest in inter-annual variation contributes greatly to the reason for the current study.

Revisions:

The reviewer is exactly correct. "Seasonal flow variation" has been replaced with "flows and their variation," which is even more precise than just deleting "seasonal."

TWDB — Required change number 3:

4th sentence, 1st paragraph, 1st page. The phrase "two flood components" should be modified to "two high-flow components." As described by TCEQ et al. (2008), high flow pulses are "in-channel, short duration, high flows." Within channel events are not generally considered "floods".

Revisions: Thanks again to the careful reviewer. The sentence has been fixed.

TWDB — Required change number 4:

The 5th sentence, 1st paragraph, 1st page is a very serious mischaracterization of the TIFP and should be modified extensively. The approach of the TIFP is not "essentially statistical" by any means! As described by TCEQ et al. (2008), "An ecosystem approach also requires the Texas Instream Flow Program to focus on essential ecological processes. Instream flow recommendations will be in the form of flow regimes containing several components. Because they occur over a range of flows, essential riverine ecosystem processes cannot be preserved by a single 'minimum' flow rate. … Based on the results of technical studies, the instream flow program will identify a set of flow components that support important processes. In general, there should be some correspondence between instream flow recommendations and historical hydrologic patters for a sub-basin." The driving factors behind instream flow recommendations for TIFP will be flow ecology relationships such as the relationship between flow rate and water temperature, dissolved oxygen, fish habitat, mussel habitat, sediment transport, oxbow connectivity, or riparian productivity. Statistical analysis of historical flow data may inform recommendations, but is not the essential basis for the choice. For example, a technical study may be used to determine flows necessary to provide for good reproduction and recruitment of a species (for example a fish species such as blue sucker or a riparian tree species such as bald cypress). Because they are long lived such species do not require suitable reproduction and recruitment flows every year for their long term survival. But they do require suitable flows every few years. An analysis of the hydrologic variability that these species have experienced historical (through a study of gaged flow records) may provide insight as to how often suitable flows related to these species (identified by the technical studies) should be provided. It is not the intent of TIFP to make recommendations that result in essential ecological process occurring more often than they have historically (nor is it clear that doing so would result in an enhanced ecological condition). Because of this concern, there is an interest in understanding the natural occurrence of various flow conditions within Texas. This study was motivated by that interest.

Revisions:

This first section was overly long and dwelled too much on environmental flows, in comparison to the other motivations for the study, so in revision it has been substantially reduced. The offending phrase has been excised.

TWDB — Suggested changes:

In order to make the document readily understandable to the largest audience with varying backgrounds and expertise, please consider writing for a lower reading level. The latest versions of Microsoft Word provide Readability Tools that may assist the author in doing so. For example, on page 10, last paragraph, the 6th sentence is worded as follows:

> "In most practical situations, the elevation of the water surface parameterizes flow in the channel, through relations from open-channel hydraulics, and also serves as an indicator variable."

This could be rewritten as

> "Typically, the elevation of the water surface (or 'stage') at a stream gage is used to estimate flow in the channel, using a rating curve developed for that location."

The first sentence has a Flesch-Kincaid Grade Level of 22.8 (i.e. post doc). The second, alternative rendering has a Flesch-Kincaid Grade Level of 16.4 (i.e. college grad). If the author could communicate to a broader audience, it would enhance the value of the final product.

Response:

This was probably a poor example of using the MS Word Readership Tools, because the revised (F-K 16.4) sentence has a different meaning than the (F-K 22.8) sentence it was offered to replace.

Generally, my approach to writing a report is to first identify the target reader, and conceive a profile of that reader's knowledge and interests, then I try to write to that level. There may be need for, on the one hand, technical diversions to elaborate details for a minority of interested readers, or, on the other hand, introductory or summary sections for readers less prepared than the target readership. In this case, the target readership was the technical staff of the Texas Water Development Board, who requested the study, and similar engineering or scientific staff at river authorities, other state agencies, and academia that may have an interest in the problem. I believe the report level is generally suitable for this readership. However, I have tried to introduce and define all important terminology to accommodate an uninitiated reader, though this is certainly redundant for the target readership.

I do agree that this is a relatively narrow readership, and the report could be modified to "communicate to a broader audience", though it's not clear that this would "enhance the value of the final product" as its intended use is to support hydroclimatological work of TWDB. Such modifications, however, would entail major revisions and substantial new text, and most of the present text would be relegated to technical appendices.

An executive summary has been provided in the revision. This is an extended abstract with selected graphics, and is directed toward a more general reader.

TWDB — Suggested changes:

In Chapter 4, your choice to use the period of 1945-2011 for the test cases comes on the heels of your statement in chapter 3 that "In Texas, Figure 9, the changing 30-year normals indicate a trend of increasing precipitation and possibly a shift in seasonality." In light of the trend picked up on in chapter 3, it might be useful to explain your choice of period in chapter 4. As you say in chapter 3, there are "philosophical issues lurking beneath the selection of a baseline period."

Response:

The demonstration cases did not extend to performing baseline-period statistics such as averaging, but are confined to an examination of the actual time series of the candidate indices. In this case, the selection of a period of analysis is mainly dictated by the need to represent a sufficient range of hydroclimatology. Because the Drought of the Fifties is the drought of record for most of Texas, it *had* to be included in the analysis period. The pluvials and intense droughts of the twenty-first century needed to be represented as well. Thus a period from 1950-2010 was more or less mandated. I extended it back to 1945 just to display an additional five years. I would like to have included the Dust Bowl years as well, but this would have overly constrained the choice of streamflow gauges.

There is of course a baseline-period implicit in the "calibration" of the SPI and the Palmer indices. For simplicity, use of the NCDC products is recommended, so in the demonstration cases, the NCDC-computed indices are used with whatever baselines NCDC employs. This does not affect the evaluation of the time series of the index for responsiveness of the index to moisture conditions, which is the main point of the demonstration. Indeed, it demonstrates how the present NCDC-computed indices — those that are presently used in the TWDB's drought monitoring activities and products provided to the Texas Drought Preparedness Council — would respond under past extremes of moisture.