Comments to the Author:

The paper was reviewed by six reviewers to which the authors have responded and extended the text. Removing the overly positive and negative reviews, leaves 4 balanced ones of which one called for rejection and three for major revisions and a second round of review.

A general undertone of the reviews was a concern about the methodology and the interpretation in your work. The responses did not address the essence of some of these concerns. Before we proceed any further with this paper I would like therefore to ask that the authors respond very clearly to the following 5 key questions:

(1) The water balance equation dS/dt = P + I – E – O (Eq. 4) is applied to the basin and the anomalies of the different components are quantified by the ratio C in Eq. 6. The 4 hydrological components are listed (magnitude and anomaly) for the 9 studied droughts in Table 2 and the ratio C in Table 3. In Table 2, the term dS/dt = P + I – E – O does not add up the number reported there (nor does the anomaly) (I tested the first drought only). Furthermore, the ratio C reported in Table 3 cannot be reconstructed from the authors Eq. 6 in the text. Although the terms do add up to 1 (as they should), their magnitude is computed in some other way. Please explain.

Response: Thank you very much for the comments. We re-examined our data processing sheets, compared with the numbers in previous Tables 2& 3, and confirmed that we made mistakes in the tables. In this revision, the mistakes were corrected.

Furthermore, in cooperation with comment (5), we found it more rational (in consistent to drought severity) to calculate the ratio Cx using the integrated dS/dt in time. So does it for precipitation, evapotranspiration, lake inflow and outflow. In the revision, we expanded our description to detail the Cx calculation (line 257-282) and modified Table 3 with the re-calculated ratios.

(2) The relative contributions of each hydrological component in Table 3 as reported indicate high variability between droughts. A key statement and conclusion of the authors is that “overall for all drought cases, 49% of the total water anomaly of the water budget came from reduced inflow, 24% from increased outflow and 18% from decreased precipitation”. I imagine this was computed from the total integrated hydrological component contributions of all droughts. How (see (1) above)? Consider that each individual drought had very different contributions, so how meaningful is this average contribution? Is it accurate and correct to summarize the conclusions on the contributions of each term to droughts in this way? Please explain.

Response: As you stated, the ratios were computed from the total integrated hydrological component contributions of all droughts. In combination with comment (5), the Cx calculation was replaced with the integrated dS/dt (also P, E, I and O), now described in the revised manuscript (line 257-282). In regard to the conclusive statements, we modified it by adding “hydroclimatic contributions to each individual drought varied from one to another, but the inflow appeared to be dominant” (line 522-524) into the text. In regard to the meanings of long-term average, we added “in many cases, quantification of individual hydro-climatic contributions to droughts is of high concern in decadal or longer time scales, for example, in climate change related studies” (line 283-285). We hope the revision could satisfy your requirement. It would be highly appreciated if you could offer us a more suitable expression. Thank you very much.

(3) Connected to the above question, I iterate one of the reviewer’s statements that you have not

answered properly: the contribution of the hydrological components during a drought are a sort of contribution to drought maintenance, not to the actual emergence (start) of a drought. In other words, a lack of precipitation in the months preceding a drought may be more important for the drought begin and nor **(not?)** precipitation deficiency during a drought, which may not be reflected at all in your analysis. Is this correct? Please explain.

Response: First, we need to make it clear whether the precipitation here refers to precipitation of the basin or precipitation of the lake region. If it is the basin precipitation, then it should at first contribute to lake inflows and then to the lake through the inflows. Therefore, the effects of the basin precipitation deficiency are included in the change of inflows. If it refers to the precipitation of the lake region, as we have indicated in the text, it played less important role compared to inflow and outflows, even in the drought periods.

Second, as we stated in the text (line 215-219), a negative SLI indicates the lake stage is lower than the normal, but not all the negatives can be classified into a drought event. Only when SLI deviates away from the normal by more than one standard deviation (SLI<-1), an event can be established (McKee et al., 1995). That is to say, a lack of precipitation may lead the lake toward dry, but not absolutely to incur a drought. In the case a drought occurs, the drought initializes when SLI becomes negative (McKee et al., 1995). In general, the negative SLI comes from the anomaly of (P-E+I-O) including the role of precipitation at the initialization time.

Definitely, meteorological droughts (precipitation deficiency) are usually ahead of hydrological droughts in time, which is stated in many literatures. The existing statements are generally qualitative. Quantitative relationships between meteorological droughts and other droughts deserve comprehensive investigation in a physics sense. There remain many questions need to be answered related to the relationships. For example, if precipitation deficiency can lead to hydrological droughts (drought emergence)? If can, when? how? To what extent? The present study quantifies lake droughts, which can serve a basis to study the questions next step.

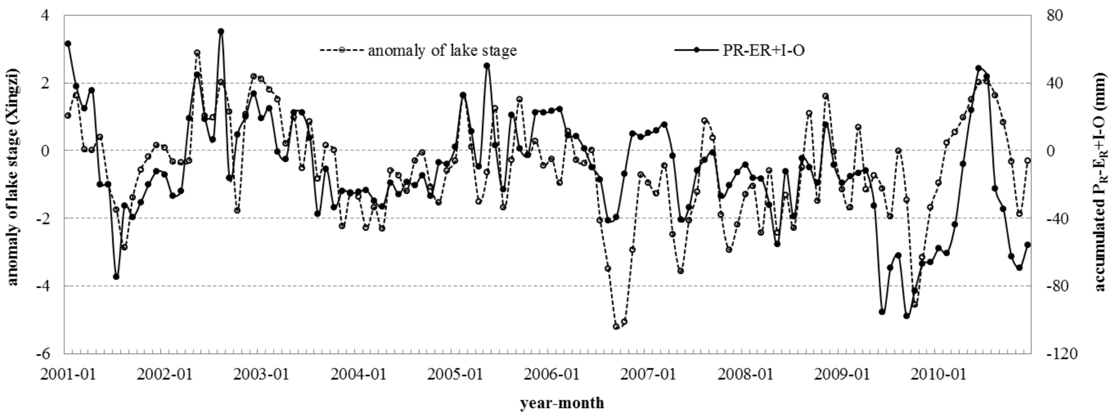
(4) It is stated that mean annual evapotranspiration is computed from the "difference between precipitation and streamflow". Which streamflow? for the entire basin or inividual subbasins? If evapotranspiration for the Lake Basin is estimated from streamflow data of the subbasins, then the relation in Fig 4b is trivial, even if it is only for individual droughts. Please explain.

Response: The multi-year mean of annual evapotranspiration was calculated from the difference between annual precipitation and discharge, respectively for the lake region and the entire lake basin in 1961-2010. The MULTI-YEAR AVERAGED monthly evapotranspiration (12-month) was then obtained from the annual value distributed with a monthly weighting factor. The monthly evapotranspiration values severs as the baseline to quantify anomaly, but not used in time-varying water balance calculation. Indeed, monthly evapotranspiration in Fig 4b comes from MODIS data varying with month and year (12-month*10-year), definitely different from the multi-year averaged values. Fig.4b is therefore an independent and direct assessment of the performance of the water balance equation for the basin.

(5) The study is based on identifying droughts from the monthly lake level data L, but it analyzes the contributions of the individual hydrological components from the general water balance equation $dS/dt = P + I – E – O$. On the average, the water balance equation is closed because mean annual E is chosen such that mean $dS/dt = 0$ (see (4) above). Throughout the paper there is no

direct assessment of the performance of the simple water balance equation for the basin. Comparing lake level data (dL/dt) and dS/dt would clearly give~~n~~ an indication how applicable the model/data are to study the posed problem, even though the lake level-volume rating curve may be complicated. Right now there is a disconnection between using L to define lake droughts and then other data (precipitation, streamflow) to define the water balance in those drought periods. In other words, if you defined droughts from the storage in the system S by integrating dS/dt in time, would you come up with the same droughts as you did with L? If not, how meaningful are then the drough period analyses of the hydrological components. Please explain.

Response: Upon the comments, the integrated dS/dt and the anomalies of lake stage and their variations for 2001-2010 are compared as shown in the following figure. As expected, they are generally in consistence and similar results can be anticipated if use of the integrated dS/dt. Notably, use of the integrated dS/dt for drought identification may be impractical. dS is general not measurable and it has to be calculated from (P-E+I-O), but the time series P+I-E-O data are often unavailable, sometimes with problems in data quality. Moreover, if without long-term data it would be hard to establish a reliable hydrologic baseline for drought quantification. Thus, use of lake stage data is preferable in quantifying drought. Anyway, the constructive comment helps us to improve the study. Thank you very much.



Hydroclimatological influences on recently increased droughts in China's largest freshwater lake

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Abstract

Lake droughts are the consequences of climatic, hydrologic and anthropogenic influences. Quantification of the droughts and clarification of contributions from the individual influences are essential for understanding drought features and their causation structure, which is highly valuable for policymakers to make effective adaption, especially under the changing climate situations. This study examines Poyang Lake, the China's largest freshwater lake, which has been undergoing drastic hydrological alternation in recent decade. Standardized lake stage is used to identify and quantify the lake droughts, and hydroclimatic contributions are determined with water budget analysis, in which absolute deficiency is defined in reference to normal hydrologic condition. Our analyses demonstrate that in the recent decade the lake droughts worsened in terms of duration, frequency, intensity and severity. Hydroclimatic contributions to each individual drought varied from one to another and the overall contribution to the decadal lake droughts came from decreased inflow (46%), increased outflow (28%), and reduced precipitation (17%) and increased evapotranspiration (9%) at the lake region. In the decreased inflow, 71% resulted from reduced precipitation and 29% from increased evapotranspiration over the Poyang Lake Basin. The increased outflow was attributable to the weakened blocking effects of the Yangtze River, in which the Three Gorges Dam (TGD) established upstream contributed less to the droughts. The TGD impoundments were not responsible for the increased drought events, but they may have intensified the droughts and changed the frequency of classified droughts. However, the TGD contribution was limited in comparison with the hydroclimatic influences. Hence, the recently increased droughts were the hydroclimatic consequences, with less important contribution from anthropogenic influences. The findings provide an example of intensified lake droughts, and offer an insightful view into lake droughts under the hydroclimatic and anthropogenic influences. It should be valuable for improving our understanding of lake droughts, and for promoting effective climate adaptation and water resources management practices.

Keywords: droughts, climate change, hydroclimatology, freshwater lake, water resources management

1. Introduction

A drought is a temporary lack of water caused by abnormal climate or environmental influences, among other factors (*Kallis*, 2008; *Mishra and Singh*, 2010; *and references therein*). There are meteorological droughts (abnormal precipitation deficits), hydrological droughts (abnormal streamflow, groundwater, or lake deficits), agricultural droughts (abnormal soil moisture deficits), ecological droughts (abnormal water deficits causing stress on ecosystems), and socio-economic droughts (abnormal failures of water supply to meet economic and social demands) (*Tallaksen and van Lanen*, 2004; *Kallis*, 2008; *Mishra and Singh*, 2010). The drought phenomena may have different temporal features and causation structure (*Kallis*, 2008; *Mishra and Singh*, 2010). It is anticipated that droughts would likely increase owing to global climate change (*Kallis*, 2008; *Mishra and Singh*, 2010).

Hydrological droughts occur when land water decreases significantly below its normal conditions, represented by low water levels in streams, lakes, reservoirs and groundwater as well (*Nalbantis and Tsakiris*, 2009; *Keskin and Sorman*, 2010). Streamflow droughts may occur with basin-scale precipitation deficiency and/or excessive evapotranspiration (*Zelenhasic and Salvai*, 1987; *Tallaksen et al.*, 1997; *Kingston et al.*, 2013). In addition to local precipitation and evaporation, lake droughts involve other hydrological components, including inflows from streams surrounding the lake and outflows out of the lake. Hence, lake droughts can be more complicated than streamflow droughts in causation structure. Furthermore, both inflows and outflows may be affected by human activities, for example, groundwater pumping, reservoir construction, or land cover change (*Wilcox et al.*, 2010). Therefore, lake droughts are the consequences of combined climatic, hydrologic and anthropogenic influences. In a disciplinary sense, in contrast to floods that have been given a great deal of attention in hydrology, droughts are not yet comprehensively understood (*Kallis*, 2008; *Mishra and Singh*, 2011). Quantification of lake droughts and clarification of contributions from individual influences are essential for understanding drought features and their causation structure, which is highly valuable for policymakers to make effective adaption, especially under the changing climate situations. Place-based drought analysis is a starting point towards integrated theories of drought (*Kallis*, 2008).

Poyang Lake is the China's largest freshwater lake, which has been undergoing drastic hydrological alterations in recent decade (*Jiao*, 2009; *Finlayson et al.*, 2010;

Hervé et al., 2011; *Liu et al.*, 2013; *Zhang et al.*, 2014). The lake is located at the south bank of the Yangtze River, which is a humid monsoon climatic region. Although the region historically experiences significant floods (*Shankman and Liang*, 2003; *Shankman et al.*, 2006), severe lake droughts have occurred frequently in the recent decade, resulting in tremendous hydrological, biological, ecological and economic consequences (*Feng et al.*, 2012; *Environment News Service*, 2012; *Wu and Liu*, 2014). Because the lake is the primary part of the well-known Poyang Lake Wetland and the lake region serves as an important food base for China, the frequently occurred lake droughts have also received increasing international attention (*Jiao*, 2009; *Finlayson et al.*, 2010; *Liu et al.*, 2011; *Environment News Service*, 2012; *The Ramsar Convention*, 2012; *Zhang et al.*, 2012, 2014).

Since lake droughts usually exhibit as an abnormal decline in lake stage or lake size, a number of studies documented the decline feature in Poyang Lake and its controlling factors (*Guo et al.*, 2012; *Zhang et al.*, 2012; *Liu et al.*, 2013; *Lai et al.*, 2014a; *Zhang et al.*, 2014). Feng et al. (2012) used 250-m satellite images and reported that the lake size had a decreasing trend between 2000 and 2010. Liu et al. (2013) revealed an abrupt decrease in the lake size in 2006, dominant in October and November. Zhang et al. (2014) demonstrated that the lake stage fell to the lowest during the 2000s compared to previous decades, in particular significantly lower in autumn recession periods. Since Poyang Lake receives inflows from its surrounding basin and discharges into the Yangtze River via a narrow outlet at the Hukou (Figure 1), the strong lake-river interaction makes it complex to separate relative impacts of the inflow and outflow on the lake stage (*Hu et al*., 2007; *Lai et al*., 2014a). Zhang et al. (2014) employed a hydrodynamic model for Poyang Lake for the separation and declared that the lake decline in the 2000s was primarily ascribed to the weakened blocking effect of the Yangtze River. Compared to climate variability on the lake basin, modifications to the Yangtze River flows have had a much greater influence on the seasonal (September-October) dryness of the lake (*Zhang et al.*, 2014). The modification was largely attributable to the operation of the Three Gorges Dam (TGD), established upstream of the Yangtze River in 2003. Water impoundments of the TGD incurred water level drops with an average estimate of 2 m at the outlet of Poyang Lake in mid-September to November for 2003-2008 (*Guo et al*. 2012; *Zhang et al.* 2012). Alternatively, Lai et al. (2013) developed a hydrodynamic model for the middle Yangtze River region (CHAM-Yangtze), in which coupled both Poyang Lake and the Yangtze River to account for the lake-river interactions explicitly. They demonstrated that the lake stage was more sensitive to the alternation in lake inflow compared to the same discharge modification in the Yangtze River (*Lai et al*., 2014a). The recent extremely low water levels in the Yangtze River resulted mainly from remarkable decline in inflows to the River, rather than solely from the TGD impoundments (*Lai et al*., 2014b). Indeed, these studies highlighted the complexity of the multiple influences on Poyang Lake's decline in the complex basin-lake-river system.

Definitely, drought differs from low water level and persistent dryness. Water level can be low in seasonal dry seasons, but it does not necessarily constitute a drought (*Smakhtin*, 2001). Persistent dryness refers to water decrease in a long run, equivalent to decreasing trend or shrinkage, which is usually unrecoverable in a short time (*Zhang et al.*, 2012). Droughts are complex events that have a recurrent feature, and it may occur in any season lasting several months or longer (*Todd et al.*, 2013). Feng et al. (2012) used 250-m satellite images to delineate inundation area for quantifying the drought severity of Poyang Lake in 2011. Linear regression analysis showed that the drought was primarily due to low basin-scale precipitation, rather than mis-balanced discharge between the Lake and the Yangtze River with TGD impoundments. Very recently, Wu and Liu (2014) used satellite-delineated inundation area to quantify two lake droughts occurred in 2006 and 2011. Statistical comparison indicated that the 2006 drought was mainly attributable to abnormal decrease of water flow in the Yangtze River and the 2011 drought was due to combined influences of the Poyang Lake Basin and the Yangtze River. Although the studies specified two extreme drought events, neither of them offered a full explanation to the frequently occurred droughts in Poyang Lake.

In principle, drought identification, quantification or characterization with a consistent standard is a prerequisite for drought analysis. However, few studies have comprehensively quantified and addressed the Poyang Lake droughts in the 2000s. The current understanding of the Lake's decline in autumn cannot provide a complete explanation to the lake droughts spanning non-autumn seasons. It remains unknown to what extent the climatic, hydrologic and anthropogenic influences have contributed to the lake droughts, which is one of key issues identified for developing integrated, interdisciplinary theories of droughts (*Kallis*, 2008). Especially for practice, clarification of the multiple influences on the recently increased droughts is essential to effective prevention of the droughts.

Complicated causality of lake droughts requires robust approach for determining the contributions from multiple influences. Analogous to standardized precipitation index, this study utilizes standardized lake index to quantify lake droughts (Section 2). With the principle of lake water balance, it proposes to define an absolute deficiency for each water component and determine their contributions to lake droughts (Section 2). The approach is applicable to basin-scale water balance, quantifying regional hydroclimatic influences on lake inflow, and subsequently on lake droughts (Section 2). Poyang Lake droughts are examined with the proposed approach, in combination with five-decade hydroclimatic data including latest satellite products (Section 3). The drought features in the 2000s and its mechanism are addressed subsequently (Section 4). Our findings should be valuable for improving our understanding of lake droughts under the changing climate situations and be useful for local water resources management and climate change adaptation.

2. Methodology

Drought has a multi-faceted nature. Major attributes of a drought event involve initiation, termination, duration, severity, magnitude, and intensity, as well as spatial extent for the case of meteorological or agricultural droughts (*Yevjevich*, 1967; *Dracup et al.*, 1980; *Wilhite and Glantz*, 1985; *McKee et al.*, 1993; *Mishra and Singh*, 2010; *Spinoni et al.*, 2014). Drought initiation time is the starting of the drought.

Termination time is the end when the drought no longer persists. Drought duration is the period between the initiation and the termination (*Yevjevich*, 1967; *Mishra and Singh*, 2010). Drought severity is the total, cumulative water deficient for the duration. Drought magnitude is a derivative of drought severity, defined as the average water deficit in the drought period (*Dracup et al.*, 1980; *Wilhite and Glantz*, 1985). Drought intensity usually refers to the largest departure from the normal conditions (*McKee et al.*, 1993; *Spinoni et al.*, 2014). For a given historical period, drought description also includes drought frequency, which refers to the number of drought events occurred (*Mishra and Singh*, 2010; *Spinoni et al.*, 2014).

2.1 Quantification of Lake Droughts

For quantitative analysis, various indices have been proposed to characterize the complex features of droughts (*Dracup et al.*, 1980; *Keyantash and Dracup*, 2002; *Mishra and Singh*, 2010; 2011). Among the numerous indices, standardized precipitation index (SPI) is commonly used (*McKee et al.*, 1993; *Mishra and Singh*, 2010). It is a normalized dimensionless index, defined as the difference of precipitation from the mean divided by the standard deviation for a given period, in which a gamma distribution is generally fitted to the long-term precipitation records for each calendar month to account for seasonal differences (*McKee et al.*, 1993). The SPI is simple but capable of quantifying drought features, and has been recently recommended by the World Meteorological Organization (WMO) to characterize meteorological droughts (*Hays et al.*, 2011). Despite it was proposed to quantify precipitation deficiency, the SPI methodology has been applied in a similar manner to other hydroclimatic variables, for example, streamflow discharge, soil moisture, reservoir storage, and groundwater level (*McKee et al.*, 1993; *Sheffield et al.*, 2004; *Vicente-Serrano and López-Moreno,* 2005; *Mendicino et al.*, 2008; *Shukla and Wood,* 2008).

In the case of lake drought, it can be described with lake stage, lake area or water storage. Because the time series data of lake area or water storage are generally unavailable, lake stage is feasible for analysis. Analogue to SPI, standardized lake index (SLI) is described as follows

$$\mathrm{SLI}_{ij} = \frac{L_{ij} - \overline{L}_j}{\sigma_j}, \qquad (1)$$

where L_{ij} is the monthly average lake stage (unit in m) of year i and month j (j = 1, 2, …, 12), normalized with a gamma distribution (*McKee et al.*, 1993). $\overline{L}_j$ is the multi-year mean of monthly average stage for month j, and σ_j is one standard deviation (S.D.) of monthly average stage for month j. Since SLI uses $\overline{L}_j$ and σ_j, both of which are monthly dependent, it thus removes seasonal differences in lake stage.

A drought event is discernible with SLI. While a negative SLI indicates the lake stage is lower than the normal, not all the negatives can be classified into a drought event. Only when SLI deviates away from the normal by more than one standard deviation (SLI<-1), an event can be established (*McKee et al.*, 1995). Furthermore, a drought initializes when SLI becomes negative and terminates before SLI becomes positive in SLI time series (*McKee et al.*, 1995). The initialization and the termination time yield drought duration (unit in day, month or year). Once all the drought events are identified, drought frequency can be determined for a given period.

In accordance with McKee et al. (1993)'s definition of SPI, SLI represents a departure of lake stage from its normal conditions. The departure corresponds to a probability of drought intensity, highly valuable to drought risk analysis. Namely, SLI=-1 denotes an occurrence probability of 15.9% (*Lloyd-Hughes and Saunders*, 2002). Furthermore, SLI can be either positive or negative. The positive (negative) value indicates lake stage higher (lower) than the normal condition for the period. The negative value quantifies the drought intensity. For an individual drought event, its lowest SLI value indicates the intensity of the event (*McKee et al.*, 1993; *Spinoni et al.*, 2014). Accordingly, a drought event can be classified into four categories with its lowest SLI: extreme drought (-∞, -2.0], severe drought (-2.0, -1.5], moderate drought (-1.5, -1.0] and mild drought (-1.0, 0.0) (*Dracup et al.*, 1980; *McKee et al.*, 1993).

In addition, drought severity of a drought event is calculated as follows

$$\mathrm{Severity} = \sum_{k=m}^{k=n} \mathrm{SPI}_k \,, \tag{2}$$

where m denotes the initialization time of a drought and n represents the termination time (*Keyantash and Dracup*, 2002; *Mishra and Singh*, 2010). Drought magnitude is then calculated as (*Keyantash and Dracup*, 2002)

$$\mathrm{Magnitude} = \mathrm{Severity}/\mathrm{duration}\,. \tag{3}$$

2.2 Contribution of Water Deficiency to Lake Droughts

A lake drought results directly from an abnormal change in lake water budget. Thus, water budget analysis is essential to clarify the drought causes. A general water balance for lake in a period can be described as follows

$$\Delta_L = P_L - E_L + G_L + I_L - O_L\,, \tag{4}$$

where Δ_L is the lake water budget (mm or m^3), P_L is the precipitation (mm or m^3), E_L is the lake evaporation (mm or m^3), G_L is the groundwater net inflow to the lake (mm or m^3), I_L is the inflow (mm or m^3) and O_L is the outflow (mm or m^3).

Once the water budget appears abnormal, it suffers from the anomalies of some or all the water components, namely, low precipitation, high evapotranspiration, low inflow and/or high outflow. At the monthly scale, for a water component X, being Δ_L, P_L, E_L, G_L, I_L or O_L, its anomaly is described as follows

$$X_{a,ij} = X_{ij} - \overline{X}_j\,, \tag{5}$$

where $X_{a,ij}$ denotes the anomaly (a) of the water component (mm or m^3) for year i and month j. $\overline{X}_j$ is the multi-year mean of X_{ij}. Notably, Eq. (5) defines an absolute water deficiency from its normal amount, different from Eq. (1) that defines a relative deficiency for drought identification. The equation offers a baseline to quantify contributions from individual water components to a drought.

During a drought event, the anomaly of lake water storage (S) in month j ($m < j \le n$) results from the consecutive anomalies of the lake water budget, which can be described as

$$S_{a,j} = \sum_{k=m}^{k=j} \Delta_{a,k}\,. \tag{6a}$$

The individual contribution from each water component to the water deficiency of the lake water storage is quantifiable with a ratio defined as follows

$$C_{X_j} = \frac{\sum_{k=m}^{k=j} X_{a,k}}{\sum_{k=m}^{k=j} \Delta_{a,k}}, \tag{7a}$$

where C_{Xj} denotes the contribution, the numerator is the sum of the monthly anomalies of the water component from month m to j. $\sum_{k=m}^{k=j} \Delta_{a,k}$ is generally negative, but $\sum_{k=m}^{k=j} X_{a,k}$ may vary with hydroclimatic conditions. For example, precipitation deficiency leads to a negative $\sum_{k=m}^{k=j} X_{a,k}$ value and produces a positive C_{Xj}. Low evapotranspiration lessens water deficiency and generates a negative C_{Xj}. Therefore, C_{Xj} can be either positive or negative. Notably, Eq.(7a) requires all the involved water components be independent from each other, which is the general case for P_L, E_L, G_L and I_L, but not for O_L. O_L is the output water, largely dependent on the sum of (P_L-E_L+G_L+I_L). To eliminate the dependence in quantifying the individual contributions with the equation, $\overline{O}_k$ is replaced with $(P_{Lk} - E_{Lk} + G_{Lk} + I_{Lk})$ for $O_{a,k} = O_k - \overline{O}_k$ which is included in Eq.(7a). This is also applicable for Eq.(7b) described subsequently.

For a drought event, in accordance with the concept of drought severity (Eq.(2)), the total water deficiency of lake water storage from the initialization time to the termination can be described as follows

$$\sum_{k=m}^{k=n} S_{a,k} = \sum_{k=m}^{k=n} \sum_{k=m}^{k=j} \Delta_{a,kk} \,. \tag{6b}$$

Then, the individual contribution from each water component to the total water deficiency for a drought event can be described as

$$C_X = \frac{\sum_{k=m}^{k=n}\sum_{k=m}^{k=j} X_{a,kk}}{\sum_{k=m}^{k=n}\sum_{k=m}^{k=j} \Delta_{a,kk}} . \tag{7b}$$

In many cases, quantification of individual hydro-climatic contributions to droughts is of high concern in decadal or longer time scales, for example, in climate change related studies (*Todd et al., 2013; Joetzjer et al., 2013; Van Lanen et al., 2013*). In this case, their individual contributions can be determined from Eq. (7b), given the total integrated hydrological components for the period.

2.3 Contribution of Basin-scale Hydroclimatic Influences on Lake Droughts

In addition to quantification of water deficiency in inputs and outputs to lake, it is natural to trace the causes of inflow deficiency for complete understanding of hydroclimatic influences on lake droughts. Lake inflow originates from precipitation in its surrounding basin. Given the water balance for the basin in a period, lake inflow is described as

$$I_L = P_B - E_B + \Delta_B , \tag{8}$$

where P_B is the precipitation (mm or m^3), E_B is the evapotranspiration (mm or m^3), and Δ_B is the change of water storage (mm or m^3), including soil moisture and groundwater in the basin.

In practice, there are often areas ungauged downstream from hydrological stations. In this case, the lake inflow includes two parts, one from gauged areas and another from ungauged areas. It has

$$I_U = P_U - E_U + \Delta_U , \tag{9a}$$

$$I_G = P_G - E_G + \Delta_G , \tag{9b}$$

where the subscript G represents the components for the gauged areas and the subscript U denotes that for the ungauged areas.

In combination with Eq. (4) and Eq. (9a), the lake water budget can be expressed as

$$\Delta_L = \underbrace{P_L - E_L + G_L}_{lake} + \underbrace{P_U - E_U + \Delta_U}_{ungauged\_area} + I_G - O_L \,. \qquad (10a)$$

Or in parallel,

$$\Delta_L = \underbrace{P_R - E_R + \Delta_R}_{lake\_region} + I_G - O_L \,. \qquad (10b)$$

where the subscript R represents the components for the lake region. It shows that the lake change involves water budgets in the lake and the ungauged areas, in addition to gauged inflow and outflow.

Further incorporated with Eq. (9b), the lake water budget can be expressed as

$$\Delta_L = \underbrace{P_L - E_L + G_L}_{lake} + \underbrace{P_U - E_U + \Delta_U}_{ungauged\_area} + \underbrace{P_G - E_G + \Delta_G}_{gauged\_area} - O_L \,. \qquad (11a)$$

Or in parallel,

$$\Delta_L = \underbrace{P_R - E_R + \Delta_R}_{lake\_region} + \underbrace{P_B - E_B + \Delta_B}_{lake\_basin} - O_L \,. \qquad (11b)$$

Clearly, the lake water change is a combined result of precipitation, evapotranspiration, soil moisture, groundwater, and outflow.

As done with Eq. (5), the water anomaly of each component in Eq. (10) and Eq. (11) can be defined, respectively. As done with Eq. (6), their contributions to the abnormal change of the lake water can be determined for water balance at different spatial scales, namely, the lake, the lake region, and the basin. The multi-scale determination is useful to trace regional hydroclimatic influences on the lake droughts.

3. Study Area and Data Processing

3.1 Study Area and Data

Poyang Lake has a maximum area of 3,860 km^2 with an average depth of 8 m at the lake stage of 22 m (*Shankman et al*., 2003). It varies remarkably from several thousand km^2 in summer to less than one thousand km^2 in winter (*Liu et al.*, 2013). There are five stations (Kangshan, Tangyin, Duchang, Xingzi, and Hukou) to measure lake stage across the lake from the south to the north (Figure 1d). It is located at the northern part of the Poyang Lake Basin, a sub-basin of the Yangtze River Basin of China (Figure 1a). The lake water flows out into the Yangtze River via the Hukou outlet. The lake water principally comes from five major river systems including Xiushui, Ganjiang, Fuhe, Raohe and Xinjiang. Seven hydrological control stations (Qiujin, Wanjiabu, Waizhou, Lijiadu, Meigang, Dufengkeng, and Shizhenjie) are located downstream to measure the discharge of the five rivers (Figure 1b). The lake region (Figure 1c) downstream from the stations is ungauged, with an area of 23,089 km^2, approximately six times of the maximum lake size. The Poyang Lake Basin has an area of 162,225 km^2 and belongs to a humid subtropical climate zone with an annual mean surface air temperature of 17.5℃ and an annual precipitation of 1,635.9 mm for the years 1960-2010 (*Liu et al*., 2012). Forestlands, agricultural fields, grasslands, bare-lands and water surfaces are the dominant land cover types (*Liu et al*., 2012).

Daily lake stage data from five hydrological stations and daily discharge data from seven control stations were obtained from the Hydrological Bureau of Poyang Lake. Lake stage data from Xingzi and Hukou were available for 1961-2010, but the data from other three stations were unavailable for 2009-2010. Daily discharge data for the Hukou outlet are available from the Hydrological Bureau of the Yangtze River Water Resources Commission. Daily precipitation data from 73 national weather stations within the Poyang Lake Basin are available from the National Meteorological Information Center of China for 1961-2010. Regional evapotranspiration data were extracted from the latest satellite products (MOD16) of the Moderate resolution Imaging Spectroradiometer (MODIS) (http://www.ntsg.umt.edu/project/mod16) (*Mu et al*., 2011) for the lake region and the lake basin in 2000-2010. In addition, the lake stage at Hukou is available for the case without the TGD for 2006-2010, which is the output of the CHAM-Yangtze model (*Lai et al*., 2013).

3.2 Drought Quantification

To identify lake droughts, SLI was calculated with Eq. (1) from monthly lake stage. Notably, it is approximate 110 km from the north to the south of Poyang Lake requiring a best representative of the lake conditions. Among five stations to measure the lake stage, the SLI values of Xingzi station had the highest correlation with that calculated from averaged lake stage of the five stations using all the available data for 1960-2008 (y=0.9953x, R^2=0.9901, p<0.0001). Thus, the station was selected for drought quantification.

Drought initialization, termination, duration, intensity, severity, and frequency were subsequently determined from the SLI values of Xingzi, with the criteria described in Section 2.1. Finally, all the lake droughts were identified and classified into extreme, severe, or moderate drought (*McKee et al*., 1993).

In addition, for the case without the TGD impoundments, the lake stage at Xingzi was estimated from its highly correlated relationship with Hukou, $y = 0.9594x+0.8034$ (*Min and Zhan*, 2013), for 2006-2010. Consequently, the SLI values of Xingzi were re-calculated for the case without TGD. It serves as a reference to evaluate the TGD effect on the lake droughts.

3.3 Water Budget Analysis

To quantify hydroclimatic influences on lake droughts, water budget analysis was designed at multi-spatial scales, preferably at the lake, the lake region and the lake basin (Figure 1). At the lake scale, water components include precipitation, evaporation, groundwater net inflow, inflows from gauged and ungauaged areas, and outflow (Eq. (4)). It has several difficulties to perform water balance analysis with a high accuracy for the lake. First, evaporation data are unavailable for the lake in monthly time series. Second, the lake has a remarkable surface variation, which significantly regulates wetland evapotranspiration (*Zhao and Liu*, 2014). Third, there are many small rivers and brooks downstream from the hydrological control stations. It is impractical to measure all of the surface runoff into Poyang Lake. Given the hydrological data, the Poyang Lake region is thus the minimum closure entity directly available for water budget analysis. For complete understanding of climatic, hydrologic and anthropogenic influences on lake droughts, water budget analysis should be performed for the lake basin, with a focus on the causes of inflow deficiency. Besides, the boundary effect of the Yangtze River is taken into consideration to account for the anomaly of lake outflow.

Specification of normal hydrologic condition is a prerequisite for determining water deficiency. First, precipitation data were grouped for the Poyang lake region and the lake basin. Multi-year mean of monthly precipitation was obtained from the observation data for 1961-2010. Second, multi-year mean of monthly discharge was calculated from the data for inflows and outflow in the period. Third, since evapotranspiration data prior to 2000 was unavailable, the multi-year mean of annual evapotranspiration was calculated from the difference between annual precipitation and discharge, respectively for the lake region and the entire lake basin in 1961-2010. The multi-year mean of monthly evapotranspiration was then obtained from the annual value distributed with a monthly weighting factor calculated from the MOD16 time series, with an assumption that the annual variability is relatively similar for 1961-2010.

Once the normal hydrologic condition was established, the water deficiency of a water component and its contribution to lake droughts were determined with Eq. (5) and Eq. (6). It was also applied to water budget for the lake region (Eq.(10b)) and the lake basin (Eq.(11b)), respectively. For the basin, a one-month lag was determined with correlation analysis between peak rainfall and peak discharge, and it was applied to account for the peak difference (*Senay et al.*, 2011; *Liu et al.*, 2013). In addition, there are three points addressed here. First, Δ_R in Eq. (10b) for the lake region is only 1.3% of the total water balance and is neglected in the present study (*Wan and*

Xu, 2010; *Zhang et al*., 2014). Second, Δ_G in Eq. (9b) for the gauged lake basin is generally unavailable. According to Feng and Liu (2014), it is roughly 5% of the total water balance and is neglected here. Third, the MOD16 datasets provide one-decade monthly evapotranspiration, making it feasible to use the independent observations for water budget analysis in time series. While the datasets have been extensively evaluated and applied worldwide, namely Australia, Brazil, Asia, and the United States (*Loarie et al*., 2011; *Kim et al*., 2012; *Velpuri et al*., 2013; *Wang et al*., 2014), our recent assessment showed that it had an error of approximate 10% for the study area (*Wu et al*., 2013). The error, together with the neglected Δ_R and Δ_G, may introduce uncertainties. Given that water anomaly (Eq. (5)) and relative contribution (Eq. (7b)) are used, the uncertainties would not be enlarged but minimized in water balance calculation and water budget analysis.

It should be emphasized that, for the sake of addressing water contribution at the lake region and the lake basin, the water amounts (unit in m^3) of all the water components were normalized to equivalent water height (unit in mm) of the whole basin (unit in km^2). Besides, statistical approaches were adopted in the present analysis (*Lomax*, 2001), in which paired F-test (and T-test) were used to examine the variance (and mean) difference between the statistics for 1961-2000 and that for 2001-2010.

4. Results and discussion

4.1 Poyang Lake Droughts in Recent Decade

Figure 2(a) illustrates the SLI variation for Poyang Lake in the recent decade.

The negative values prevail over the positive, indicating the dry phase dominates the lake for the period. There occurred three extreme, two severe and four moderate droughts, according to the drought classification criteria (*McKee et al.*, 1993). Among the nine cases, three droughts started in spring, two in summer and four in autumn (Table 1). Drought duration varied from 2 to 13 months with a mean of 6.2 months and one standard deviation (S.D.) of 4.1 months, which demonstrated that the lake droughts could occur in any month. Drought intensity ranged from -1.03 to -3.03 with a mean of -1.79±0.70. The top three lowest SLI values were -3.03, -2.66 and -2.01, corresponding to possibilities of 0.12%, 0.39% and 2.22%, respectively, for each occurrence. Drought severity varied from -1.39 to -20.22 with a mean of -7.16±6.41. More specifically, in the category of 'extreme drought', the drought event that ranked first in both intensity and severity occurred from July 2006-July 2007, lasting 13 months. The 2006 drought was addressed in Feng et al. (2012) and Wu and Liu (2014) in terms of inundated area, whereas the present study quantified its probability of occurrence and revealed that the drought lasted longer than the previous reports. The second most severe drought event emerged in September 2009-January 2010, persisting 5 months. The third most severe drought took place from October 2007-August 2008, lasting 11 months. The two droughts in the 'severe drought' category spanned 6 and 10 months, respectively. The four droughts in the 'moderate drought' category lasted 2~4 months. It appears that a drought with a lower SLI is usually more severe and lasts for a longer time.

In comparison to the years 1961-2000, the lake droughts changed in terms of duration, frequency, intensity and severity in the most recent decade (Figure 2(b)). On average, drought duration extended from 5.6 to 6.2 months. Drought frequency increased from 6.0 to 9.0 events per decade. Drought intensity intensified from -1.38 to -1.79, and drought severity increased from -5.02 to -7.16. In regard to the intensification, further analysis revealed that the moderate drought events increased. The severe droughts decreased, but the extreme droughts increased from 0.5 to 3.0 events per decade (Figure 2(c)). Overall, the lake droughts have worsened in terms of duration, frequency, intensity and severity over the last decade.

4.2 Hydroclimatic Change at Poyang Lake Region

Normal variation of water components is a baseline for quantitative analysis of drought occurrence as an abnormal change. Figure 3(a) shows the multi-year mean of monthly precipitation ($\mathbf{P_R}$) and evapotranspiration ($\mathbf{E_R}$) for the lake region, lake inflow (**I**) from five major rivers and outflow (**O**) into the Yangtze River. The monthly precipitation varied with a peak in June followed by a sharp decrease. Inflow had a similar seasonal pattern. Outflow had the maximum value in June and the minimum in January. The maximum evapotranspiration appeared in August and the minimum in December. From a perspective of water balance, the water budget was positive from January to June with a peak in June. It became negative from July to December, and the minimum value appeared in October. These results indicate a shift in water budget from a surplus phase in the first half of the year to a deficit phase in the second half of the year. The deficit phase is a part of the normal hydrologic condition, and thus it does not necessarily mean a drought occurrence. Furthermore, in annual water budget, the equivalent water supply from the local precipitation was 312.0 mm, and that from inflow was 714.4 mm, more than two times of the local precipitation. The water loss from the local evapotranspiration was 118.3 mm, and that from outflow was 908.1 mm, approximate 7.7 times of the local evapotranspiration. It demonstrated that the inflow and outflow were much higher than the local precipitation and evapotranspiration. This finding implies the dominant role of hydrologic components over meteorological components in regulating Poyang Lake.

Lake droughts occur when abnormal change appears in the water budget. Table 2 lists the water components for the lake region during the periods of lake droughts. The water budgets ($\mathbf{P_R}$-$\mathbf{E_R}$+**I**-**O**) were -67.1, -12.6 and -69.3 mm for three extreme lake droughts, -71.6 and 27.7 mm for two severe lake droughts, and 72.0, 5.4, 18.5, and -66.1 mm for four moderate lake droughts. In sum, the budget was negative (deficit) for five cases and positive (surplus) for four cases. Despite the positive water budgets, the large negative anomalies of $\mathbf{P_R}$-$\mathbf{E_R}$+**I** for the cases illustrated that the lake water income ($\mathbf{P_R}$-$\mathbf{E_R}$+**I**) was exceptionally lower than normal. The low water income resulted from largely decreased inflow and precipitation, as well as increased evapotranspiration. The positive water budgets were attributed to the water surplus period in the first half of the year. In this sense, the definition of drought is a water anomaly referenced to a normal state of either water surplus or deficit phase. It indicated that a drought occurrence was more closely related to the water deficiency (negative anomaly) of water budget than the net water budget. For example, the net water budget did not show statistically significant relationships with drought intensity or drought severity. On the contrary, the total water anomaly of a drought event showed a significant relationship with drought severity (x) (y =22.949x+3.8772, n = 9, R^2=0.7513, p < 0.01) (Figure 3(b)). In general, the water budget analysis highlighted the importance of water deficiency in reference to a normal condition of either water surplus or deficit phase.

While a lake drought is generated from the anomaly of water budget, it is not clear to what extent each water component has contributed to the anomaly. Clarification of the contribution is helpful for understanding drought causes and will be useful for preventing droughts under the changing climate. Table 3 shows the ratios of the total water anomaly of a component to that of the water budget for each event (Eq. (6)). For the three extreme droughts, outflow and inflow played the most important role. For example, inflow was 210.7 mm lower, local precipitation was 100.9 mm lower and evapotranspiration was 44.8 mm higher than normal for the most severe drought (July 2006-July 2007). Outflow was 296.3 mm lower as a result of reduced inputs to the lake. For the second most severe drought (July 2007-August 2008), precipitation was 39.0 mm lower and evapotranspiration was 31.6 mm higher. Inflow was 141.9 mm lower, contributing to 23% of the anomaly. For the third most severe drought (September 2009-January 2010), precipitation was 12.0 mm lower and evapotranspiration was 11.2 mm higher. The reduced inflow contributed to 25% of the anomaly and the increased outflow contributed to 56% for the period. In addition to the positive contribution, a water component may contribute negatively. For example, there are two negative cases of outflow occurring mainly in a water surplus period

(Table 3). The negative contribution implies less outflow in reference to water income ($\mathbf{P_R-E_R+I}$), which is consistent with the water budget ($\mathbf{P_R-E_R+I-O}$). Clearly, hydroclimatic contributions to each individual drought varied from one to another, but the inflow appeared to be dominant. On average, for all the drought cases, 46% of the total water anomaly of the water budget came from the reduced inflow, 28% from increased outflow, 17% from decreased precipitation and 9% from increased evapotranspiration at the lake region. Since inflow reduction is the major contribution to the water anomaly in the lake region, it is vital to trace how precipitation and evapotranspiration have changed at the basin scale in the recent decade.

4.3 Hydroclimatic Change at Poyang Lake Basin

Likewise, prior to performing a water budget analysis, it is valuable to clarify the normal hydrologic condition. Generally, precipitation ($\mathbf{P_B}$) and evapotranspiration ($\mathbf{E_B}$) had similar seasonal patterns in the basin as its counterpart in the lake region (Figure 4(a)). Monthly precipitation varied seasonably with a peak in June, followed by peaks in May and April. Major precipitation appeared in the first half of the year. Monthly evapotranspiration was generally less than precipitation and its top three highest values appeared from June-August. Monthly outflow was approximately half of precipitation with a similar seasonal pattern. Consequently, the monthly water budget was positive (surplus) from December-June and negative (deficit) from July-November. The highest water surpluses appeared in March, April and May, and the lowest water deficits in July, August and September. On an annual scale, outflow occupies approximately 55% of precipitation, 10% higher than evapotranspiration, which is one of the climate features of this humid subtropical region.

Concerning water budget for a drought event, it was -252.2, -79.4 and -95.8 mm for the three extreme droughts, -251.9 and 152.0 mm for the two severe droughts, and 136.3, 21.3, -9.5 and -65.1 mm for the moderate droughts (Table 2). For six negative cases, the water budget featured less precipitation (negative anomaly) and more evapotranspiration (positive anomaly). For three positive cases, $\mathbf{P_B}$-$\mathbf{E_B}$ had large negative anomalies over 100 mm, but the water budgets became positive due to the largely reduced outflow. For the three extreme droughts, $\mathbf{P_B}$-$\mathbf{E_B}$ was much lower than the normal, suggesting that meteorological droughts have made significant effects on the drought formation. For all the cases, the water anomalies of $\mathbf{P_B}$ and $\mathbf{P_B}$-$\mathbf{E_B}$ had positive relationships with drought severity, which was consistent with the water budget for the lake region. Nevertheless, the basin-scale precipitation is the most important water source to the lake, as confirmed by a correlation between $\mathbf{P_B}$-$\mathbf{E_B}$ (x) and **I** (y) (y=0.9958x, R^2=0.8667, n=10, p<0.005) (Figure 4(b)). The $\mathbf{P_B}$-$\mathbf{E_B}$-to-**I** difference was -28.3 mm, approximately 10% of **I**, in agreement with our previous study (*Wu et al.*, 2013). The high correlation and relatively small difference also confirmed the suitability of the satellite evapotranspiration data for the study area.

While the inflow reduction generated from combined hydroclimatic change, precipitation and evapotranspiration may have made different contributions to the lake droughts. For example, precipitation was 248.0 mm lower and evapotranspiration was 72.5 mm higher than the normal, and they produced inflow 152.8 mm lower from October 2007-August 2008. The decreased precipitation contributed 83% and the increased evapotranspiration contributed 17% to the inflow deficiency, respectively. In contrast, precipitation was 23.3 mm lower and evapotranspiration was 55.1 mm higher, and they generated inflow that was 21.1 mm lower than normal from June-November 2001. The corresponding contribution was 39% for precipitation and 61% for evapotranspiration to the inflow deficiency. For all nine-drought cases, 71% of reduced inflow came from decreased precipitation and 29% came from increased evapotranspiration as a whole (Table 3). These results demonstrated the dominant role of precipitation in reducing the inflow to Poyang Lake, despite it might vary with different drought cases.

4.4 Mechanisms Accounting for Recent Lake Droughts

The above sections detail the lake droughts as abnormal phenomena and the hydroclimatic contribution to individual drought events. Yet, it remains unclear why the droughts strengthened in the recent decade, and whether the droughts resulted from a long-term change of hydroclimatic influences or a seasonal combination of these influences.

Figure 5(a) shows the accumulated anomalies of water budget from 2001-2010. At the lake region, the water budget $\mathbf{(P_R-E_R+I-O)}$ declined from mid-2002 to a low value in September 2009, and then increased yet remained in a negative phase. Obviously, the decrease in the water budget is a hydroclimatic setting for the recent drought increase. The water deficits involve local precipitation and evapotranspiration, lake inflow and outflow, but each of these components has different courses. The accumulated $\mathbf{P_R}$ showed a decreasing trend after mid-2003. The accumulated $\mathbf{E_R}$

increased gradually but steadily, which was consistent with the rapid increase of surface temperature in the Poyang Lake Basin since 1998 (*Liu et al*., 2012). The $\mathbf{E_R}$ exceeded the $\mathbf{P_R}$ after April 2010, exhibiting an increasing effect on the water budget. Comparatively, the accumulated **I** or **O** had a relatively large variation, consistent with their dominance over $\mathbf{P_R}$ and $\mathbf{E_R}$ at seasonal scale. They displayed similar behaviors with a peak in spring 2003, and then declined by the end of 2009. In the entire period, precipitation decreased by 5%, evapotranspiration increased by 19%, inflow declined by 5% and outflow declined by 4%, accounting for the negative water budget in the lake region.

Figure 5(b) displays seasonal variation of the water budget during 2001-2010. In comparison to 1961-2000, the water surplus reduced for the first half of the year, and the water deficit increased for the second half of the year except for August and November. The large reduced surplus includes March and June, and the enhanced deficit includes July and September. The reduced surplus and the enhanced deficit would increase the possibility of drought occurrence and intensify the drought intensity. In the enlarged water deficit, the $\mathbf{P_R}$-$\mathbf{E_R}$ and the **I-O** contributed to 43% and 57%, respectively. In the **I-O** deficit, inflow decreased but outflow increased. Usually, the outflow decreases with reduced inflow and $\mathbf{P_R}$-$\mathbf{E_R}$. Since the Yangtze River serves as a boundary condition of Poyang Lake, the increased outflow is generally a result of weakened blocking effects of the lake-River interactions (*Guo et al*., 2012; *Zhang et al*., 2012; *Lai et al*., 2014).

The weakened effects involve climate change in the upper reaches of the Yangtze

River, and water impoundments of the TGD (*Guo et al*. 2012). Routinely, the TGD impoundment begins in mid-September and spans one to two months. Among all the drought events, none occurred during exactly the same time span. Accordingly, the TGD impoundments should not be responsible for the increased drought events. However, the impoundments lowered the lake stage at the Hukou outlet by 1~2 meters for September-October (*Guo et al*. 2012; *Zhang et al*., 2012, 2014). Our analysis indicated that the impoundments led to a change in SLI from -2.70 to -3.03 for the extreme drought in July 2006-July 2007, and from -1.81 to -2.66 for the extreme drought in September 2009-Januray 2010. The change has two implications. First, the droughts were intensified with the TGD impoundments. Second, a severe drought ($-2.0<SLI\le-1.5$) was intensified to an extreme drought ($-\infty<SLI\le-2.0$), which changed the frequency of classified droughts. This is a reasonable explanation for the decrease in the number of severe droughts but increase in extreme droughts (Figure 2(c)). Furthermore, according to the latest lake storage curve described in Tan et al. (2013), the lowered lake stages would result in a water loss of 7.1×10^8 m^3 and 24.1×10^8 m^3 for each event, respectively. The losses occupied 1.5% and 19.1% of the total anomalies of outflow, corresponding to 0.2% and 9.9% of the contribution to each drought event. In comparison with the hydroclimatic influences, the TGD contributions were rather limited.

In general, the recently increased droughts were principally attributed to decreased inflow, increased outflow, and reduced local precipitation and increased evapotranspiration at the lake region. The findings lay in several strengths or novelties in the present study. First, use of satellite-retrieved evapotranspiration data makes it possible to analyze drought causes from a perspective of water budget with independent measures of major water components. Given measurement errors are quality controlled, the independent observations are more faithful than model simulation that is susceptible to model uncertainty and empirical parameterization. Indeed, the MOD16 products demonstrated its effectiveness in water balance calculation and subsequent water budget analysis. Second, in addition to drought quantification, absolute deficiency was defined for water component and water budget in reference to normal hydrologic state. Individual hydroclimatic contributions were isolated from the total anomalies of water budget, and the drought causation structure was subsequently distinguished. The quantification approach is straightforward and applicable to separate hydroclimatic influences to droughts, a key issue identified for developing integrated theories of droughts (*Kallis*, 2008). Third, it is the first time to quantify all the drought events and their causation structure in Poyang Lake. Most existing studies did not explicitly quantify the droughts but focused on low water levels mainly in autumn seasons. A few studies addressed 1-2 extreme droughts with statistical regression analysis (*Feng et al.*, 2012; *Wu and Liu*, 2014), yet lack of systematic water balance analysis of the droughts and their decadal features. The present study completed drought quantification, water budget analysis, isolation of hydroclimatic contributions, and clarification of causation structure for the recently increased droughts in Poyang Lake. The complete assessments demonstrated that the droughts were the hydroclimatic consequences, with less important contribution from the TGD influences. Yet, it should be noted that the present study did not address some potential influences, for example, land cover/use change, agricultural water use, soil moisture variation and vegetation dynamics. These factors can significantly affect the hydrological processes at seasonal and annual scales, and subsequently affect lake stage and droughts, which should be taken into consideration in future study.

5. Conclusions

This paper used standardized lake stage to identify and quantify droughts illustrated with the case in Poyang Lake for all seasons. From a perspective of water budget, it defined an absolute deficiency for water component and water budget in reference to normal hydrologic condition to determine hydroclimatic contributions to lake droughts. Given five-decade hydroclimatic observation data and the latest satellite products, water budget analysis was operational and performed in the study area.

Our analyses demonstrated that the lake droughts had strengthened in the recent decade, in terms of duration, frequency, intensity and severity. The overall contribution to the lake droughts came from decreased inflow (46%), increased outflow (28%), and reduced precipitation (17%) and increased evapotranspiration (9%) at the lake region. In the decreased inflow, 71% resulted from basin-scale precipitation and 29% from basin-scale evapotranspiration. The recently increased droughts were principally ascribed to hydroclimatic change, specified with decreased precipitation, increased evapotranspiration and reduced inflow. The TGD impoundments were not responsible for the increased drought events, but they did intensify the droughts and change the frequency of classified droughts. However, the TGD contribution was limited, compared with the hydroclimatic influences.

Overall, the findings provide an example of intensified lake droughts, and offer an insightful analysis of the droughts in the changing climate and anthropogenic influences. It should be valuable for improving our understanding of droughts, thus benefiting to develop integrated theories on the subject. This study is also useful for the effective promotion of water resource management and climate change adaptation.

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Figure Captions

Figure 1. Geographic location of Poyang Lake, China. The lake is principally fed by a five river systems of the Poyang Lake Basin. Lake water flows into the Yangtze River via a sole outlet at the Hukou. Jiujiang is located 25 km upstream of the Hukou on the Yangtze River. The Three Gorges Dam (TGD) is upstream of the river.

Figure 2. Poyang Lake droughts in 2001-2010. (a) Variation in the standardized lake stage index (SLI). (b) Drought duration, frequency, intensity and severity, and (c) Drought frequency for moderate, severe and extreme droughts, compared to 1961-2000.

Figure 3. (a) Multi-year mean of monthly precipitation ($\mathbf{P_R}$) and evapotranspiration ($\mathbf{E_R}$) for the Poyang Lake region, lake inflow (**I**) from five major rivers of the Poyang Lake Basin, and outflow (**O**) into the Yangtze River. All the water amounts are normalized to equivalent water height of the whole Poyang Lake Basin. (b) The relationship between drought severity and total water anomaly of ($\mathbf{P_R}$-$\mathbf{E_R}$+**I**) of each

event for nine cases of Poyang Lake droughts.

Figure 4. (a) Multi-year mean of monthly precipitation ($\mathbf{P_B}$) and evapotranspiration ($\mathbf{E_B}$) for the Poyang Lake Basin, and outflow (**O**) into the Yangtze River. (b) The relationship between $\mathbf{P_B}$-$\mathbf{E_B}$ and inflow for nine cases of Poyang Lake droughts. All the water amounts are normalized to equivalent water height of the whole Poyang Lake Basin.

Figure 5. (a) Accumulated anomaly of water components and (b) water budget at the Poyang Lake region for 2001-2010 compared to 1961-2000. All the water amounts are normalized to equivalent water height of the whole Poyang Lake Basin.