We thank the reviewer for detailed review and very useful comments. The following is our reply to the comments raised.

1) At site KY, surface water $\delta^{18}O$ is mainly determined by the oceanic end-member. The estimated f_{CFW} is only 0 to 5% for the last 7 ka. Furthermore, the binary mixing model contains numerous assumptions (ex. since the KSSW $\delta^{18}O$ record for the last 7 ka does not exist, the KSSW$\delta^{18}O$ values are deduced from a KSW record assuming a constant offset between KSSW and KSW; TSW$\delta^{18}O$ is estimated from the result of one core of which only two intervals were dated by ^{14}C; The mixing proportion of KSW, KSSW and TSW is supposed to be constant for the whole studied period; The cave temperature is also supposed to be constant, and under this condition the$\delta^{18}O$ values of the freshwater end-member are calculated). The authors consider uncertainty related to Mg/Ca-SST estimate to evaluate the error of f_{CFW}. Since each assumption adds distinct uncertainty to the f_{CFW} estimate, it is not certain that the small proportion of the river water contribution is always significant when all the uncertainty is propagated. This point should be examined.

>>$\delta^{18}O_w$ of Kuroshio Surface Water ($\delta^{18}O_{KSW}$) is estimated as $\delta^{18}O_{KSW}$ =1/3 ($\delta^{18}O_{2404}$+$\delta^{18}O_{2403}$+$\delta^{18}O_{A7}$) based on $\delta^{18}O_w$ data sets reconstructed at three sites, MD01-2403 (Lin et al., 2006), MD01-2404 (Chen et al., 2010), and A7 (Sun et al., 2005), in the Okinawa Trough. Evenly spaced $\delta^{18}O_w$ data with 0.01 ky spacing were made by resampling from original unevenly spaced $\delta^{18}O_w$ time series data using a software *AnalySeries*. Because the maximum age uncertainty for these data sets is about 300 years, three sets of 300-yr moving averaged data were prepared and used for calculating $\delta^{18}O_{KSW.}$ The difference between calculated $\delta^{18}O_{KSW}$ and individual 300-yr averaged $\delta^{18}O_{2404}$, $\delta^{18}O_{2403}$, and $\delta^{18}O_{A7}$ at the same age does not exceed ±0.26‰. Therefore, we assume the error of $\delta^{18}O_{KSW}$ estimate (ε_{KSW}), which is due mostly to spatial heterogeneity and age uncertainty, as ±0.26‰.

Although the geological record of the past $\delta^{18}O_w$ of Kuroshio Subsurface Water ($\delta^{18}O_{KSSW}$), which is defined as a water mass at 100 m water depth in the area of 122-123ºE, 24-25ºN at present, from 7 ka to present does not exist, we assume that the difference in $\delta^{18}O_w$ between KSW and KSSW has been kept small (<0.1‰) during the Holocene because of the following reason. At modern condition, the difference in $\delta^{18}O_w$ between KSW and KSSW is estimated as <0.1‰ from the salinity difference of <0.5 (Data from Japan Ocean Data Center (JODC) available at http://jdoss1.jodc.go.jp/cgi-bin/1997/bss.jp) between them during June to August (the most stratified season) and $\delta^{18}O_{KSW}$ is lower than $\delta^{18}O_{KSSw}$ owing to excess precipitation over evaporation in the surface. Inter-annual salinity variability is larger ($1\sigma = \pm 0.06$‰) in the surface in the defined area, which is mostly caused by excess precipitation over evaporation during summer, compared to at ~100 m depth, the latter is more stable within ±0.018‰ (JODC). Therefore, the surface freshening during summer is the main factor that controls the salinity (and $\delta^{18}O_w$) difference between the surface and subsurface water, and it is reasonable to consider that the $\delta^{18}O_{KSW}$ value has been always lower than that of $\delta^{18}O_{KSSW}$. On the other hand, the maximum offset during the last 7 ky is difficult to estimate. However, if we consider that vertical mixing due to winter monsoon wind is the major factor to reduce the offset between the surface and subsurface, and intensity of winter monsoon wind over the East China Sea is expected to have been stronger during the middle Holocene based on a multi-model analysis (Zhao and Harrison, 2012), it would be reasonable to assume the offset has never been larger than 0.1‰. Thus, the minimum and maximum offset between $\delta^{18}O_{KSW}$ and $\delta^{18}O_{KSSW}$ is assumed to have been 0‰ and 0.1‰. As is described above, the uncertainty in $\delta^{18}O_{KSW}$ estimation due to the spatial heterogeneity and age uncertainty is ±0.26‰ during the last 7 ky. So, the uncertainty in $\delta^{18}O_{KSSW}$ estimation should be smaller than ±0.26‰ because the modern inter-annual variability in $\delta^{18}O_{KSSW}$ (=salinity of KSSW) is less than $\delta^{18}O_{KSW}$ (=salinity of KSW). Thus we estimate the $\delta^{18}O_w$ offset between KSW and KSSW as 0.1‰+0.26‰/-0.1‰.

The end-member $\delta^{18}O_w$ of the Kuroshio Taiwan Water (KTW) was expressed as follows using $\delta^{18}O_w$ of Taiwan Strait Water ($\delta^{18}O_{TSW}$), $\delta^{18}O_{KSW}$, and $\delta^{18}O_{KSSW}$ by applying the modern mixing ratio of the water masses, when each error is expressed as ε.

$\delta^{18}O_{KTW} \pm \varepsilon_{KTW} = 1/2 \cdot (\delta^{18}O_{TSW} \pm \varepsilon_{TSW}) + 1/2 \cdot (\delta^{18}O_{KSW} \pm \varepsilon_{KSW}) + 1/4 \cdot (0.1 \pm \varepsilon_{KSSW})$

Then, a propagated error of KTW (ε_{KTW}) is expressed as follows.

$\varepsilon_{KTW} = (1/2 \cdot \varepsilon_{TSW} + 1/2 \cdot \varepsilon_{KSW} + 1/4 \cdot \varepsilon_{KSSW})$

Assuming $\delta^{18}O_w$ of core MD01-2904 as the end member of TSW, then, ±0.12‰ that is stemmed from Mg/Ca and $\delta^{18}O$ can be applied as the error of $\delta^{18}O_{TSW}$. When ε_{TSW}=0.12‰ and $\varepsilon_{KSW} = \varepsilon_{KSSW}$ =0.26‰, ε_{KTW} is calculated as 0.26‰.

The error stemmed from changes in the mixing ratio of the water masses should be smaller than ε_{KTW} of 0.26‰, because the differences among calculated $\delta^{18}O_{KTW}$, $\delta^{18}O_{TSW}$, $\delta^{18}O_{KTW}$, and $\delta^{18}O_{KSSW}$ are small as explained as follows. The difference between the calculated $\delta^{18}O_{KTW}$ and $\delta^{18}O_{KSW}$ ($\delta^{18}O_{KSSW}$) at the same time slice is always within ±0.26‰. Similarly, the difference between $\delta^{18}O_{KTW}$ and $\delta^{18}O_{TSW}$ at the same time slice is within ±0.26‰ except for the time interval between 7 and 6.8 ka when the differences between them are about 0.3‰. At 7-6.8 ka, the differences would be within ±0.26‰ if the contribution of TSW on KTW could be considered to be less than 90%. Considering the modern mixing ratio of the water masses described in the above equation, more than 90% contribution of TSW seems unlikely. Therefore, the error stemmed from changes in the mixing ratio of the water masses can be considered to be within ±0.26‰, and the possible maximum error of $\delta^{18}O_{KTW}$ is ±0.26‰.

Next, we evaluate the air temperature effect on calculating the Changjiang River freshwater contribution (f_{CFW}). While we calculated the freshwater end member with the constant cave temperature during the last 7 ka in the previous manuscript, here we consider a case that the temperature gradient exists between the 7 ka and late Holocene. The air temperature in inland China at 7-6 ka is estimated to be ~2 ºC higher than today and decreased toward the late Holocene based on a pollen assemblage record (Shi et al., 1993). Fig. 1 shows a result of the time series of f_{CFW} that is obtained by assuming that the air temperature decreases monotonously by 2 ºC from 7 ka to the present. Because the $\delta^{18}O_{CFW}$ becomes lower in this case, the obtained f_{CFW} becomes higher. However, the differences in the obtained f_{CFW} between the two cases with and without the air temperature gradient are 0.2% at most, which means that the effect of the 2º C changes in air temperature is very small. Thus, the air temperature effect does not affect the conclusion that indicates no prominent long-term trend in the obtained f_{CFW} record since the middle Holocene. Therefore, the absence of long-term decrease in $\delta^{18}O_w$ at site KY is a robust feature and not due to the changes in the end-member $\delta^{18}O_w$ but due to the absence of long-term decrease in the Changjiang freshwater flux.

2) The changes in f_{CFW} are interpreted in terms of the past EASM precipitation variability. But other factors, such as monsoonal winds, might have significant influence to the Changjiang river water advection. As authors state in modern climatological settings, stronger southerly wind could enhance the eastward extension of Changjiang diluted water, leading to higher f_{CFW} values even if Changjiang river discharge is invariable. Such alternative possibility should be discussed.

>> Jiang et al. (2008) reported that the strength of the southerly wind during summer is negatively correlated with summer precipitation in the Changjiang Basin for the last ~ 50 years. That is, when the southerly wind is stronger, the Changjiang Basin gains less precipitation. Nevertheless, the salinity around site KY has a robust negative correlation

with the flux of the Changjiang, suggesting that the influence of the southerly wind on the f_{CFW} around site KY is very small. For example, during 1956 to 1960 when the southerly wind was stronger than normal year, while the discharge of the Changjiang freshwater is lower (Jiang et al., 2008), the salinity around site KY was also higher.

3) It is not clear for me whether the centennial to sub-millennial scale variability of f_{CFW} (Fig. 8) is real and correctly estimated. The authors average and smooth $\delta^{18}O$ values of three oceanic water masses (KSW, KSSW and TSW) because the three records do not show similar variability (Fig. 6). The difference between KSW, KSSW and TSW $\delta^{18}O$ records is explained by "large analytical error, local variability of precipitation/ evaporation, or large error in $\delta^{18}O_w$ attributable to heterogeneity of the samples". Due to averaging and smoothing, centennial to sub-millennial scale variability of $\delta^{18}O$ records of the oceanic end-member is erased (Fig. 7) whereas the centennial to submillennial scale variability of the surface water $\delta^{18}O$ record at site KY is maintained. Does the high frequent variability of f_{CFW} remain even if the smoothing of the oceanic component is omitted?

>> As the reviewer #2 pointed out, the time resolution of the data sets are different among the cores. The time resolution of our data is highest (~30 yrs), while others are roughly ~100-200 yrs. To avoid a discrepancy in time resolution, we use the 300-yr averaged data sets, which are evenly spaced ahead, for calculation of f_{CFW}. We believe that 300-yr averaging is reasonable when we take into account the age uncertainties, which are at most ±150 years, among the cores. The reconstructed f_{CFW} is shown in Fig. 1, indicating submillennial scale variability of the $\delta^{18}O_w$ at site KY still exists. Although the uncertainty of the reconstructed f_{CFW} becomes larger than the previously calculated f_{CFW} because of the propagated error described above, submillennial scale variability is marginally visible.

4) I do not understand the interest of flux estimate (section 4.4) by adding further hypotheses. The flux variability (Fig. 10) and fCFW changes (Fig. 8) are virtually the same.

>> As the reviewer #2 pointed out, the flux variations are proportional to f_{CFW} changes. However, we believe it is important to show the magnitude of the flux variability because the numerical number of the flux makes it easy to compare the past record with modern variability of the Changjiang discharge that helps people to imagine how large the past variability was compared to the modern variability.

5) Except for the El Niño record by Moy et al. (2002), there is no comparison between f_{CFW} and other EASM records, time series of forcing (ex. solar insolation, solar activity) and modelling results. This lack makes difficult to evaluate the robustness of the authors' main message.

>> Taking into account the comments by the reviewer #1, we omit the comparison between our record and Moy et al. (2002). Instead, we add comparison between our record of the Changjiang freshwater discharge and speleothem $\delta^{18}O$, insolation, and modeling results shown in Fig. 1. Please see our reply to the reviewer #1's comment 6.

6) It is possible that speleothem $\delta^{18}O$ records cannot be totally explained by summer monsoonal intensity. However, modelling studies also indicate the influence of solar insolation on the EASM intensity (ex. Liu et al., 2003; Kutzbach et al., 2008). Consequently, it seems difficult to justify the different EASM intensity evaluated by this study and speleothem records only by the bias of speleothem $\delta^{18}O$ records. Indeed, the authors do not give any explanation about the absence of long-term trend of Changjiang river water discharge. Taken together, I suggest whole revision of paper including re-evaluation of propagated uncertainty of f_{CFW}, comparison with possible forcing, other reconstructed time series and modelling results, and explanation of the insensitivity to the local insolation. It is necessary to clarify the absence of local insolation effect is a local feature or a more regional trend.

>> We agree with the suggestion of reviewer #2 to give more detailed discussion about our results. In the revised manuscript, we add a section "5.2. Regional difference in timing of the Holocene optimum precipitation" in the discussion chapter, in which we compare our results with other proxy records in China and model results. It is believed that the stronger boreal summer insolation in the Northern hemisphere in the early to middle Holocene compared to today have enhanced EASM precipitation, which are supported by earlier modeling studies (e.g., Kutzbach et al., 2008). In contrast, a recent transient simulation study for the Holocene revealed the complexity of the response of the Asian summer monsoon system to the insolation change during the Holocene (Jin et al., 2014). For example, the northern area (northern China, southern Mongolia) and southern area (southwestern and southern China) of the EASM could have higher precipitation than today during the early to middle Holocene, while the central and eastern area of the EASM (middle reaches of the Yangtze River (Changjiang) Basin, Korea, and Japan) could have less precipitation then. This spatial heterogeneity is attributed to internal feedbacks within climate system, such as the air-sea interactions associated with the El Nino/Southern Oscillation and/or shift of the Inter Tropical Convergence Zone (Jin et al., 2014).

The flux of the Changjiang freshwater estimated by this study shows no apparent long-term trend from the middle Holocene to the present-day, suggesting that temporal changes in the EASM precipitation in the Changjiang Basin do not simply follow the Holocene insolation pattern that monotonically decreases. In contrast, there are evidences that suggest decline in EASM precipitation from early/middle Holocene to present-day in other EASM areas, such as northern China (Zhang et al., 2011). This difference suggests that the insolation is not simply affect the intensity of EASM precipitation but the internal feedbacks mentioned above could be also important to control the spatio-temporal pattern of EASM precipitation.

Minor or specific comments

1) The title of the paper would be modified because the reconstruction is not really quantitative taking into account the uncertainty.

>> We follow the advice and change the title as follows.

"Changes in East Asian summer monsoon precipitation during the Holocene deduced from the Changjiang (Yangtze River) freshwater flux reconstruction based on oxygen isotope mass-balance in the northern East China Sea"

2) In the introduction, the authors focus on possible bias of speleothem $\delta^{18}O$ records as indictors of the EASM intensity. In contrast, they concentrate on ENSO influence in discussion section. The manuscript should be reorganized to be consistent.

>> Thank you for the comment. We add discussion on comparison with speleothem $\delta^{18}O$ records. The absence of the long-term trend in our f_{CFW} suggests that speleothem $\delta^{18}O$ does not reflect precipitation amount in this region but other factors such as $\delta^{18}O$ of the precipitation (moisture), reflecting changes in the moisture source, $\delta^{18}O$ itself in the source, and seasonal precipitation amount. Following the reviewer #1's comment, we omit the discussion on the comparison with the ENSO record of Moy et al. (2000).

3) P. 1449, lines 14-15. Introduction. The authors state that the tight linkage between the intensity of EASM and local summer insolation on orbital timescales is based on the speleothem $\delta^{18}O$ records. This is not true because modelling studies also indicate the influence of solar insolation on the EASM intensity (ex. Liu et al., 2003; Kutzbach et al., 2008).

>>We will modify the introduction part and add citation of the modeling studies that indicate the influence of summer insolation on EASM intensity.

4) P. 1450, lines 4-7. The variation of compiled lake level records within Changjiang Basin might be compared with f_{CFW} (Fig. 8).

>> In Fig. 8 of the revised manuscript, we will compare our f_{CFW} data with insolation, speleothem $\delta^{18}O$ and the regional lake-level record to discuss the EASM precipitation in the Changjiang Basin.

The reconstructed Changjiang freshwater flux in this study does not show the long-term decreasing trend from the middle Holocene that is instead apparent in the Chinese speleothem $\delta^{18}O$ (e.g., Wang et al., 2005). An et al. (2000) argued that temporal-spatial pattern of the EASM precipitation was asynchronous across China and it could be explained by the southward retreat of the monsoon front during the Holocene. They inferred that the EASM precipitation peak was at 8-5 ka in the middle and lower reaches of the Changjiang River based on a climate model simulation, pollen, and lake-level records. However, the high precipitation peak that expected at 8-5 ka is not evident in the compiled lake-level records in the middle and lower reaches of the Changjiang River, and the lake-levels seems to have increased slightly since the middle Holocene (An et al., 2000). Therefore, at least, the view that the EASM precipitation was reduced from the middle Holocene is not supported by both our Changjiang flux record and the compiled lake-level record. Besides the lack of decreasing trend in both our Changjiang flux record and the compiled lake-level record in the middle and lower reaches of the Changjiang River by An et al. (2000) further suggests that the regional summer precipitation intensity does not seem to be the main controlling factor of the Chinese stalagmite $\delta^{18}O$. However, there is a small discrepancy between our f_{CFW} record and the compiled regional lake-level record. Namely, our f_{CFW} record does not have a long-term increasing trend from the middle Holocene, which is suggested by the compiled regional lake-level record. Possible explanation of this discrepancy is that the number of the lake-level record that were used for the compiled record may not be enough to capture the average of the precipitation/evaporation balance over the entire Changjiang Basin and some of the lake-level record may reflect local effect such as changes in hydrological connection with local rivers or changes in human activities.

5) P. 1450, line 12, "CDW". Please define this word at the first use.

>> We apologize for this. We add the definition of CDW as Changjiang Diluted Water.

6) P. 1450, lines 11-13. The calcification depth of *G. ruber* is estimated to be upper 30 m in this study. Did the author distinguish different morphotypes of *G. ruber* sensu strict and sensu lato? Since *G. ruber* (s.s.) calcifies indeed in the upper 30m but *G. ruber* (s.l.) calcifies below 30 m (Wang, 2000), only *G. ruber* (s.s) should have used in this study.

>> We only use *G. ruber* (s.s.). We add this information in the revised text.

7) P. 1456, line 7. "For core KY core," should be "For core KY,".

>> corrected.

8) P. 1457, line 2. "from11.6 to" should be "from 11.6 to".

>> corrected.

9) P. 1461, line 3. A reference (Chen et al., 2010) is missing in the reference list.

>> added.

10) P. 1464, line 24. "four data set" should be "four data sets".

>> corrected.

11) P. 1465, lines 5-16. The authors use speleothem $\delta^{18}O$ record of Heshang Cave to estimate $\delta^{18}O$ of Changjiang fresh water. They indicate that the Sanbao Cave $\delta^{18}O$ record gives consistent results. How about the estimate based on the speleothem record of Dongge cave shown in Fig. 1?

>> $\delta^{18}O$ of Dongge cave also gives consistent result in terms of long-term trend.

12) P. 1467, lines 19-29. The salinity data for the period 1985-1990 was deviated from a general trend because of the decrease in salinity of the end-member. The interest here is why the salinity of end-member declined for this period.

>> The $\delta^{18}O_w$ of the endmember KTW is controlled by local precipitation/evaporation balance. The lower salinity could be explained by higher precipitation in the Kuroshio water area or Kuroshio source area.

13) P. 1467, line 29-P. 1468, line 1. The data for period 1996-2000 is omitted due to a large annual variability. Again, why the annual variability was abnormally large?

>> There is a large interannual variability in surface salinity data for the period of 1996-2000. This is because the extremely low salinities of ~30 were observed in summer of 1998 when the severe flood event occurred. On the other hand, salinities in other years from 1996 to 2000 are more or less the same and between 33.4 and 33.8. The standard deviation of the salinity for 1996-2000 is 1.3, which exceeds 2σ (=1.2) of the ~ 50 year salinity data. The standard deviations of the salinity for the other periods are less than 0.5. Therefore, we omit the salinity data for 1996-2000, because its standard deviation exceeds 2σ and may not be representative of the average value. In 1998, precipitation over the catchment and the floodwater discharge from the upper basin did not exceed the historical maximum, but water levels in the middle basin were much higher than the historical maximum (Zong and Chen, 2000). It means that much larger freshwater drained into the East China Sea in short time and this extreme freshwater discharge might have been the cause of the extremely low salinity.

14) Fig. 6. Please indicate raw data points of *G. ruber* Mg/Ca and surface water $\delta^{18}O$ of each record in addition to running average curves and/or temporal resolution of original data sets in the figure caption. This information is helpful to judge whether different centennial to sub-millennial variability observed for each core is related to its temporal resolution. Please indicate age control points for core 2904.

>>We did not show original data points of each record because the Fig. 6 became messy in the previous manuscript. As is suggested, we add the original data points of each data set and age control points for core 2904 in the revised figure in the revised manuscript.

15) Reference. Chen et al., 2010 and Wang, 2000 are missing in the list.

>> Added in the reference list.

References

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Zong, Y., and Chen, X., The 1998 Flood on the Yangtze, China, Natural Hazards, 22, 165-184, 2000.

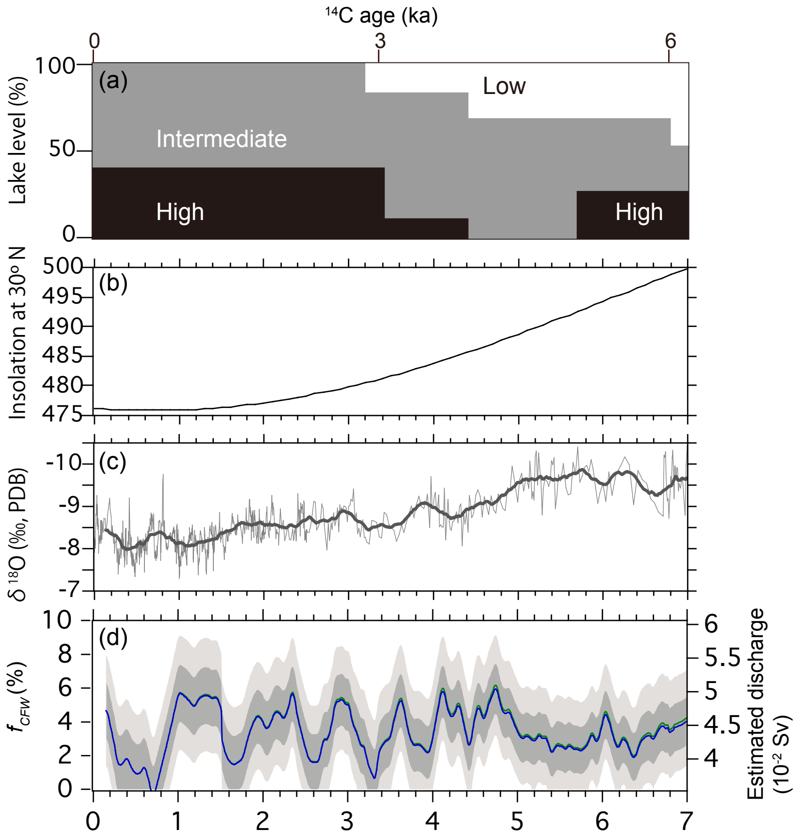

Fig. 1. (a) The regional lake-level record in the Changjiang Basin compiled by An et al. (2000) but presented in the conventional ^{14}C age. (b) Insolation on 21th June in 30º N (Lasker et al., 2004). (c) Speleothem δ^{18}O in Heshang Cave (Hu et al., 2008). (d) Relative contribution of the Changjiang freshwater (f_{CFW}) at site KY in the northern East China Sea and estimated Changjiang freshwater discharge. Light and dark gray shaded area indicate uncertainties propagated error of 0.5σ and 1σ, respectively. The green and blue lines indicate estimated f_{CFW} with the 2 ºC temperature gradient since the middle Holocene and with constant cave temperature at 17 ºC, respectively.