# Terrestrial water fluxes dominated by transpiration: Comment

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The fraction of evapotranspiration (ET) attributed to plant transpiration ($T$) is an important source of uncertainty in terrestrial water fluxes and land surface modeling (Lawrence et al. 2007, Miralles et al. 2011). Jasechko et al. (2013) used stable oxygen and hydrogen isotope ratios from 73 large lakes to investigate the relative roles of evaporation ($E$) and $T$ in ET from their catchments and to estimate global $T$/ET. Two recent contributions have addressed data issues with their work highlighting uncertainty propagation (Coenders-Gerrits et al. 2014) and global interception estimates (Coenders-Gerrits et al. 2014, Schlesinger and Jasechko 2014). We address here the underlying assumptions of the model applied by Jasechko et al. (2013). They assumed that the isotope ratios of the lake water incorporate

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spatially integrated fractionation effects of total $E$ and $T$ over the entire catchment. This assumption is invalid and leads to substantial errors both for catchment-scale and global estimates of $T$/ET.

Several processes at the catchment-scale underlying the assumption that lakes integrate watershed ET are misrepresented or overlooked in the paper. First, soil water pools undergoing ET do not completely mix with water that contributes runoff to lakes in catchments where precipitation is >ET, and are rarely recorded in catchments where precipitation is ≤ET (Kendall and McDonnell 1998, Brooks et al. 2010). The H and O isotope ratios of water in streams, shallow groundwater and lakes therefore do not necessarily carry the isotope signal of total $T$/ET from all terrestrial surfaces, despite evidence of substantial soil $E$ in the isotope signals within plant waters (Tang and Feng 2001); such $E$ signals are not apparent in the waters of many lakes even within large, arid catchments (Henderson and Shuman 2009). Second, Jasechko et al. (2013) do not include sublimation ($S$) in their analysis as an important water flux at the catchment scale or as a potentially non-fractionating isotopic process (Earman et al. 2006). Consequently, the isotope mass balance models presented are appropriate for open, non-frozen water, but do not accurately capture isotopic exchange effects in terrestrial ecosystems (Kendall and McDonnell 1998) or snow-dominated catchments. We offer new data demonstrating that lakes do not consistently record H and O isotope signals from catchment-scale $E$ and $T$ (our main point) and that $T$ is not the only non-fractionating isotopic process with significant water loss to the atmosphere (a secondary point).

Jasechko et al. (2013) extrapolated catchment-scale estimates to global $T$/ET by calculating global deuterium excess of $E$ as an average of catchment values (Jasechko et al. 2013, their Appendix S2); this method, however, assumes that the mean of their 73 catchments is an unbiased statistical representation of global water fluxes, that all relevant water fluxes are correctly estimated and that isotopic signatures are correctly interpreted. This method requires that every water flux that is important for a

**Wyoming snow, rain, groundwater, streams, and lakes**



Fig. 1. Isotopic values of lakes (orange) from two regions in Wyoming, USA, stream (blue squares), and groundwater (blue circles) show that evaporative trends (local evaporation lines, LELs) develop within lakes. In Wyoming, precipitation (crosses, gray for snow, blue for rain) and snowpack data reveal that substantial moisture loss from the snowpack through sublimation (Fig. 3) does not cause the isotopic values of the residual snowpack to diverge from the local meteoric water line, and confirms that lake inputs represent the mean of the annual range of precipitation rather than the mean of the snow-dominated surplus (which would be the case if a substantial fraction of growing season rain was transpired before water flows into the lakes). Each symbol indicates a different lake. Isotopic values are reported in per mil relative to international reference material, Vienna standard mean ocean water (VSMOW).

catchment be included in the analysis (e.g., decoupling of lake and terrestrial water in arid catchments, or snow sublimation in snow-dominated catchments). Consequently, a biased estimate for any of the 73 catchments will multiply to large errors when $T/ET$ is extrapolated to the global scale.

Our data reveal that the H and O isotope ratio values of lake water from watersheds across North America differ from the values of groundwater and stream samples collected at the same time (Figs. 1 and 2). The relationships in the data indicate that the lakes derive little of their

isotopic value from the terrestrial landscape, and thus cannot provide the hoped for information about $T$. Lake inputs cannot be differentiated from adjacent streams and groundwater, which share the isotopic values of mean annual precipitation. The lakes, but not the stream and groundwaters, include the only meaningful $E$ signal. Consequently, the related $E$ must take place within the lake from the open water and not the surrounding watershed, despite isotopic evidence of $E$ within soil waters (Fig. 2 and, e.g., Allison et al. 1983, Barnes and Allison 1988, Tang and Feng 2001). The isotopic values of soil water
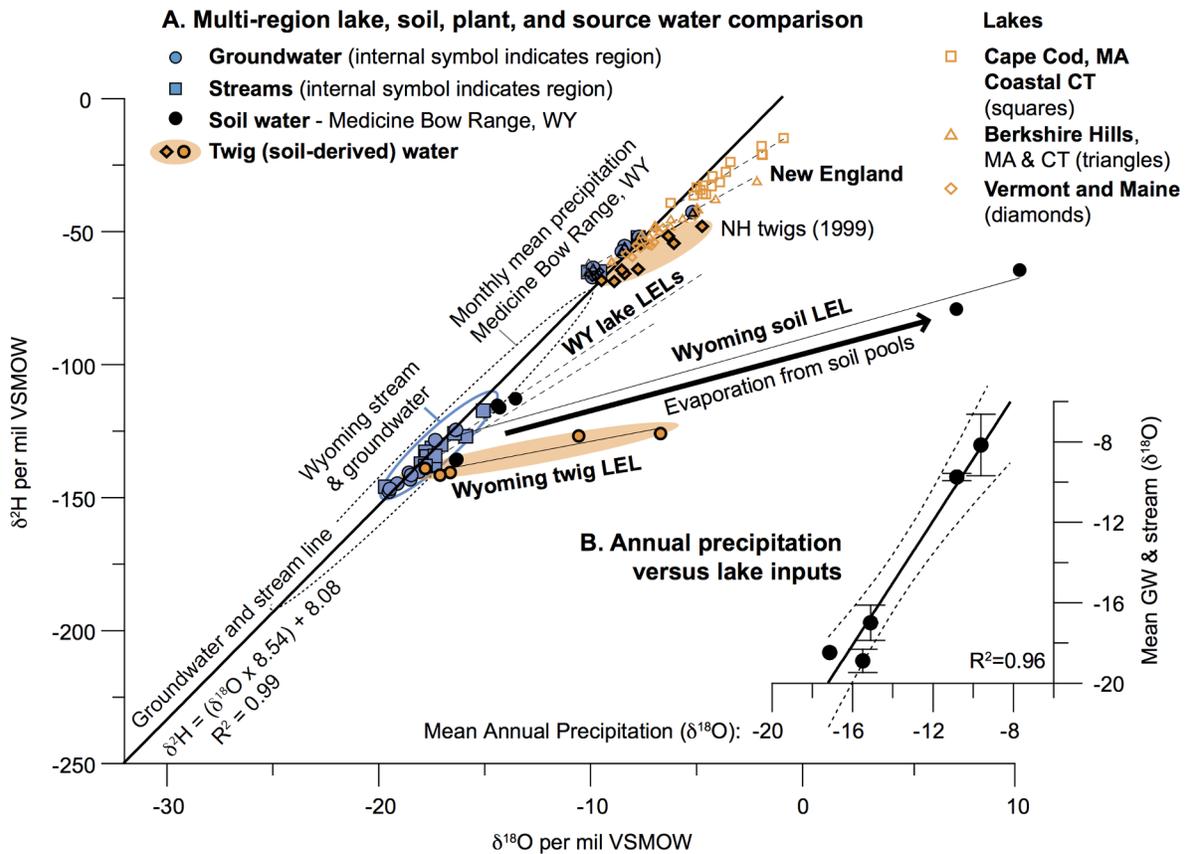
Fig. 2. Isotopic values of lakes (orange), stream (blue squares), groundwater (blue circles), soil water (black circles), and twigs (symbols in orange ovals) from five regions across North America show that different evaporative trends (local evaporation lines, LELs) develop within lakes and soils (A). Lake inputs (blue) originate from the mid-point of our rain and snow samples for Wyoming rather than from a mean of the snow data. The relationship that lake water derives from non-enriched groundwater and streamwater, holds for multiple regions from the Rocky Mountains to the northeastern United States and is consistent with little evaporative influence on the lake inputs, despite the appearance of evaporative enrichment in terrestrial soil (black circles) and plant waters (data in orange oval; New England data from Tang and Feng (2001) for comparison). The inset scatter plot (B) shows the correlation between the annual mean precipitation and the lake inputs. A dashed oval indicates the range of mean monthly precipitation values as calculated by Bowen and Revenaugh (2003). Each symbol indicates a different lake.

also tend to evolve along local evaporative lines with lower slopes than evaporation from open lake waters (Gibson et al. 2008), but mixing of these two signals is not apparent. The observation that the E signal from lake water originates from within lakes appears consistent with Fig. S1 by Jasechko et al. (2013) and statements in published literature (e.g., "Isotopic composition of lakes is controlled predominantly by variations in evaporation loss from lake and upstream lakes," in Gibson and Edwards 2002).

Two explanations may lead to the observation that E signals in soil water are not transferred by stream and groundwater to lakes, but they have different implications for lake isotopic budgets: (1) If non-fractionating processes, especially $T$, dominate within the watersheds (despite some limited $E$) as proposed by Jasechko et al. (2013), we would expect high $T$ rates during the growing season to remove water from terrestrial pools without fractionation, bias lake inputs towards isotopically light surplus (because isotopically

heavy growing-season moisture has been transpired or evaporated before deep soil infiltration), and thus differentiate input waters from mean annual precipitation (Henderson and Shuman 2009). However, such a difference is not supported by our data, which do not differentiate the isotopic signal of lake input waters from mean annual precipitation (Figs. 1 and 2). Alternatively, (2) when water rapidly flows through soil to streams it may by-pass the soil water pools, which are ultimately released to the atmosphere by ET. If ET removes most or all water from shallow soil pools, then lakes and streams will receive inputs carrying little isotopic information about upland processes. The presence of soil-derived $E$ signals in plant stem waters (Tang and Feng 2001), but not in the ground and stream waters of the same region (Fig. 2), appears consistent with this possibility. $E$ from terrestrial surfaces is thus underestimated because water flowing to lakes is decoupled from water that was influenced by $E$ in the terrestrial landscape. Furthermore, annual (Jasechko et al. 2013) rather than seasonal isotopic budgets could mask the seasonally-biased effects of $T$, if $T$ preferentially removes growing season moisture. However, whether either interpretation is correct is currently an unresolved question pending further research. Thus, it is too early to base either catchment-scale or global ET estimates on one or the other, particularly without quantifying related uncertainty.

Arid regions serve as an obvious example demonstrating one extreme point (of complete decoupling) on a continuum of decoupling between lake water and catchment-scale processes. Isotopic signals of ET within catchments in dry regions, which cover 30–40% of the terrestrial surface (Millenium Ecosystem Assessment 2005, Peel et al. 2007) and contain >350 Gt C (17–36% of global totals; Trumper et al. 2008, Trumper et al. 2009), are not integrated over their entire area and are not represented in lakes because inputs originate only from the wettest parts and flow through dry areas without much processing (Bailey 1979). In contrast, Jasechko et al. (2013) assume complete coupling and mixing of lake water and catchment-level processes; in their words: "The pathway water takes after falling as precipitation within a catchment includes mixing, evaporation (fractionation labelled) and transpiration (non-fractionation labelled), until the remaining water accumulates in a downstream lake or river. Each of these catchment processes is ultimately recorded by the isotopic composition of the lake's water." Consequently, they underestimate catchment-scale $E$ of dry regions by over-representing the wetter parts with denser vegetation that have a larger $T$/ET ratio than the drier parts of these catchments, which do not propagate their $E$ signal to lakes. Thus, they estimate $T$/ET $\sim$ 0.75 for desert catchments. To support this result they incorrectly cite as field data a modeling study from the Jornada Experimental Range in southern New Mexico (Reynolds et al. 2000). The Reynolds et al. (2000) study contained no field estimates of $T$/ET, but reported a range of simulated annual $T$/ET of 0.01–0.6 and an average annual value over the 100-year simulation period of 0.34.

Our data from monitored watersheds in the Rocky Mountains indicate that T is not the only non-fractionating isotope process (Figs. 1 and 2) with significant water loss at the catchment scale (Fig. 3). First, our data suggest that the isotopic effects of snow sublimation ($S$) are complex and may have no or a minor net influence on deuterium-excess (Fig. 1). While $S$ can lead to enrichment of $^{2}$H and $^{18}$O and changes in deuterium excess in the residual snowpack (e.g., Koeniger et al. 2008), some authors find that condensation in the snowpack can cancel the isotopic effects of $S$ completely (Moser and Stichler 1975, Friedman et al. 1991) or partially (Gustafson et al. 2010). At our research site, we do not observe an isotopic signal of $S$ with a detectable kinetic isotope effect in our snowpack or lake water samples; all snowpack, stream, and groundwater samples fall along the local meteoric water line with no change in deuterium excess (Fig. 1). This result suggests that even if interpretation (1) from the previous paragraph holds true, then $T$ and $S$ may be confounded in the isotopic signature of lake water. Second, our data suggest that $S$ can be a significant flux in the annual water budget ($\sim$1/2 of total annual water vapor flux; Fig. 3) in snow-dominated catchments. Literature supports the assumptions of Jasechko et al. (2013) that $S$ is insignificant at a global scale (e.g., Dirmeyer et al. 2006, Lawrence et al. 2007, Miralles et al. 2011). However, these
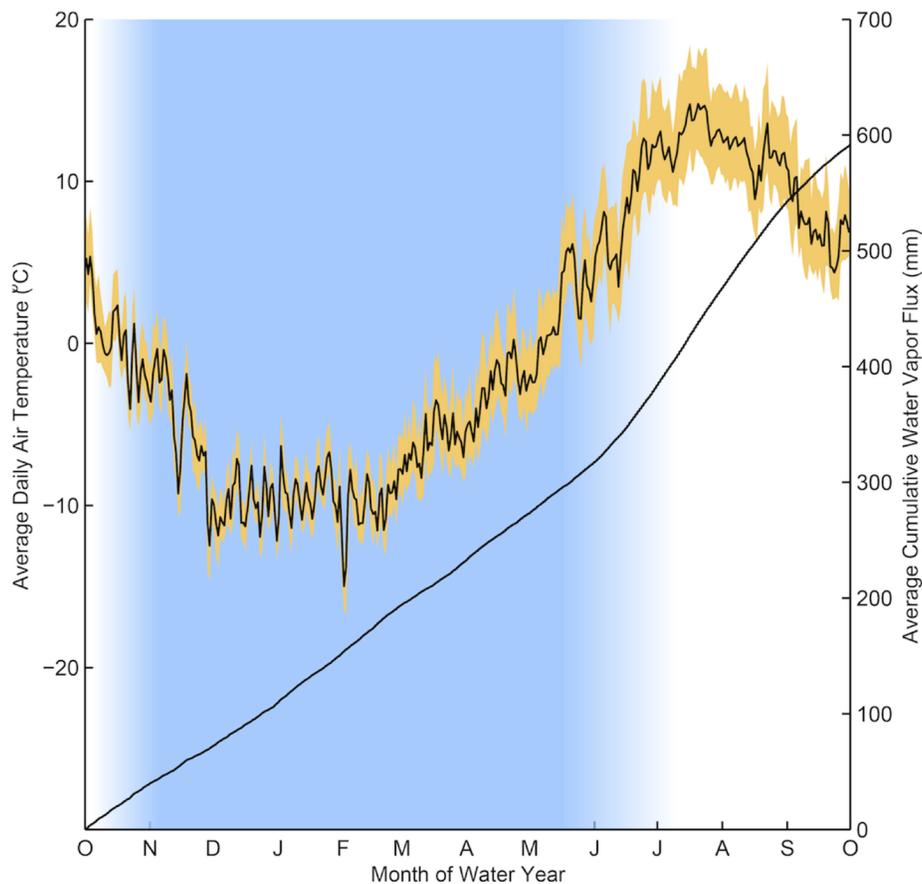
Fig. 3. Average cumulative water vapor flux (solid line) for the past eight water years at GLEES. The snowpack persists for an average of eight months per year (blue shading) while the climate is below freezing a majority of that time (line with yellow shading: average daily minimum, mean, and maximum air temperatures). Sublimation ($S$) accounts on average for $50 \pm 10\%$ (SD among years) of the total annual water vapor flux.

global $S$/ET values are based on modeling and not observations. In contrast, many observations and catchment-scale estimates, in addition to our data, clearly demonstrate a large contribution of $S$ to annual ET. The importance of $S$ increases with canopy interception of snow (Parviainen and Pomeroy 2000, Koeniger et al. 2008) and turbulent or blowing snow situations (Déry and Yau 2001, Reba et al. 2012) as well as with the ratio of snowfall to total annual precipitation. For instance, $S$ removed 4–41% of seasonal snow accumulation in a catchment of the central Rocky Mountains (Reba et al. 2012), 20–30% of snowfall events in a subalpine forest (Montesi et al. 2004), 23% of seasonal snowfall in a continental desert (Zhou et al. 2012), 20–40% of annual snowfall in boreal forests (Pomeroy et al. 1998), and 35% of

arctic annual precipitation (Liston and Sturm 2004). Even if global $S$ is much smaller than $E$ or $T$, $S$ can be significant in catchments with water inputs derived from snow. Furthermore, snow-dominated catchments contain ~384 Gt C (~19% of global totals, Trumper et al. 2009) and are widespread, i.e., almost all catchments poleward of 40–45° latitude (Barnett et al. 2005, Adam et al. 2009). Thus, most of the catchments in North America, Europe, and Asia considered by Jasechko et al. (2013) are snow-dominated and $S$ is likely an important water flux of these catchments. Because their analysis ignores $S$, but $S$ is likely high in most of their catchments, they implicitly assume that the isotopic signal of $S$ to be equal to $E$. Any observed net isotopic signal of $S$ smaller than that of $E$, e.g., our data (Fig. 1),

will cause an under-estimation of $E$ according to the methods of Jasechko et al. (2013). For instance, data from the research site at GLEES (Fig. 3) are comparable to some of their catchments in the western United States, for which Jasechko et al. (2013) estimate $T/ET \sim 0.7$–1 (their Fig. 3A). Yet we measured $S/ET \sim 0.5$; thus $T/ET$ must be $<0.5$ and likely smaller because $E > 0$. Consequently, ignoring $S$ and its unique isotope exchange processes may lead to substantial over-estimation of $T/ET$ (Jasechko et al. 2013) in areas of the world where snowmelt dominates streamflow (Barnett et al. 2005, Viviroli et al. 2007) and also in their global estimates.

Field estimates of the partitioning of ET into $T$ and $E$ are expensive and difficult, but are becoming more common with the availability of field-portable water vapor isotope analyzers. The majority of published studies are based on simulations (e.g., Reynolds et al. 2000); observational studies frequently bias towards periods of high plant activity and thus overestimate $T/ET$ over longer periods. For instance, a controlled experiment across a gradient of canopy cover measured $T/ET = 0.83$ for full canopy cover, but this value is only valid for the daytime and on small plots of 0.36 m$^2$ (Wang et al. 2010); other field measurements lasted a growing-season and measured a floodplain forest with groundwater access with $T/ET = 0.9$ (Yepez et al. 2007). The overestimation of $T/ET$ is apparent when compared to measurements of a growing-season in a similar system without groundwater access that obtained $T/ET = 0.37$–0.44 (Cavanaugh et al. 2011). Therefore, results of such studies cannot easily be extrapolated or compared to annual values and for larger geographic extents such as watersheds. There are a few multi-year field estimates that are at a scale that makes them comparable with values obtained by Jasechko et al. (2013). Oishi et al. (2010) measured sap flux to estimate $T$ and eddy covariance to estimate ET, found $T/ET = 0.54 \pm 0.03$ (mean $\pm$ 1 SD) during four years in a mature oak-hickory forest in North Carolina, USA, with annual precipitation between 934-1346 mm. Similarly, Raz Yaseef et al. (2010) found $T/ET = 0.49 \pm 0.06$ during four years in a 40-year-old pine forest in Israel with annual precipitation of $285 \pm 88$ mm. Thus, these and other observational studies support the notion that the $T/ET$ values presented by Jasechko et al.

(2013) are considerably too high.

Jasechko et al. (2013) included five validation approaches to their results. Their first approach compares their 'isotope-based with conventional open-water evaporation rates' (see their Fig. S1). As discussed above, this comparison underscores our concern that lakes do not fully integrate catchment processes because the comparison demonstrates with a high $R^2 = 0.78$ that their $E$ estimates represent within-lake open-water $E$ and carry thus only a small if any isotopic signal from the terrestrial surface. Their second validation approach uses the 'geographic similarity between compiled in situ transpiration measurements' and their estimates, i.e., they compared their values with two citations (Yakir and Wang 1996, Williams et al. 2004). As noted above, many field measurements bias towards greatest plant activity temporally and spatially and Yakir and Wang (1996) is no exception: fluxes were measured at midday for crops at the field scale and are thus likely much higher than if measured in natural vegetation during an entire year. Williams et al. (2004) measured in a surface irrigated olive orchard during 15 days and found $T/ET = 0.69$–0.86 after irrigation and $T/ET = 1$ prior to irrigation. These values can be put in perspective by a more thorough study that measured water balance in a natural forests over four years and found $T/ET = 0.49$ (see discussion above; Raz Yaseef et al. 2010). We suggest that the second validation approach is thus biased and incomplete. Their third validation approach is based on an 'agreement between $^{18}O/^{16}O$- and $^2H/^1H$-based evaporation rates using equation (3)'. Our arguments and data demonstrate that their equation 3 is inadequate because important terrestrial signals are not incorporated in either the O or H component of the lake water; it is only appropriate for open, non-frozen water (Figs. 1 and 2) but not for terrestrial ecosystems or snow-dominated catchments. Their fourth validation approach is based on their equation (4) and 'shows that transpiration accounts for 80% to 90% of terrestrial evapotranspiration'. Eq. 4 is equivalent to their Eq. 3 except that deuterium excess is used. Hence, this is not an independent validation as it is based on the same underlying model that we criticize (see also our comment to their third validation approach). Their fifth validation approach applies a global estimate of

water use efficiency (WUE) of the terrestrial biosphere to their global transpiration flux to calculate gross primary production and compares this result to other published estimates. One of the problems with this approach is that their definition of WUE is at the plant-level, which, for instance, does not account for $E$ losses from soils and is inappropriate for catchment-scale and global estimates (Niu et al. 2011). Overall, Jasechko et al. (2013) did not successfully validate their model and results (i–ii, v) and did not provide independent validation (iii–iv).

We conclude that lake water does not fully integrate all processes and areas within every catchment, and even if it did, the magnitudes of ET components, such as $T$ and $S$, are not uniquely identifiable using isotope ratio measurements. Therefore, the methods presented by Jasechko et al. (2013) lead to substantial over-estimation of $T$, by under-estimation of $E$ and/or by under-estimation of $S$, and, thus, the relative importance of $T$ to ET for individual catchments, and when extrapolated to global $T$/ET. Their method may only be suitable for those catchments where lake water represents a complete and well-mixed sample of all terrestrial surface processes and where no precipitation falls in the form of snow; however, this is not the case for much of the terrestrial surface. Consequently, the net effect of their method is a very large overestimation of the relative importance of $T$ as a percentage of total catchment and global ET.

## METHODS

Lake ($n = 90$ samples), stream ($n = 16$), and groundwater ($n = 16$) were sampled in five North American regions (Cape Cod, MA and Coastal CT; Berkshire Hills, MA and CT; VT and MA; Medicine Bow and Park Ranges, WY and CO; Northern WY) shown in Figs. 1 and 2 during 2008–2011. Lakes were sampled at the same time as associated groundwater and streams. Snowpack ($n = 67$), snowfall ($n = 9$) and rain ($n = 2$) were collected at the GLEES AmeriFlux site (Medicine Bow Range, WY) in spring 2013. Samples were placed into sealed containers with no air headspace, and transported to the lab where hydrogen and oxygen isotope measurements were performed using an off-axis integrated cavity output spectrometer (Baer et al. 2002).

Flux data were collected using the eddy covariance technique (Lee et al. 2004) at the GLEES AmeriFlux site (http://ameriflux.ornl.gov/) from 2004–2012 (Frank et al. 2014). Sublimation was defined as all vapor flux occurring with a snowpack, but limited when $CO_2$ flux from non-dormant vegetation was evident (Reba et al. 2012).

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