



Precipitation Extremes Influence Patterns and Partitioning of Evapotranspiration and Transpiration in a Deciduous Boreal Larch Forest

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1 **Precipitation extremes influence patterns and partitioning of evapotranspiration**
2 **and transpiration in a deciduous boreal larch forest**

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4 1, 2JiaLin Liu, 3FangYan Cheng, *2, 4J. William Munger, 2, 5Peng Jiang, 2Timothy G.
5 Whitby, 4, 6SiYue Chen, 4, 6WeiWen Ji, *7XiuLing Man

6
7 1Arts and Science, New York University Shanghai, 1555 Century Avenue, Shanghai,
8 200122, China;

9 2School of Engineering and Applied Sciences, Harvard University, 29 Oxford Street,
10 Cambridge, 02138, USA;

11 3School of Life Sciences, Fudan University, 2005 Songhu Road, Shanghai, 200438,
12 China;

13 4Department of Earth and Planetary Sciences, Harvard University, 20 Oxford Street,
14 Cambridge, 02138, USA;

15 5College of Urban and Environmental Sciences, Peking University, 5 Yiheyuan Road,
16 Beijing, 100871, China;

17 6School of Physics, Peking University, 5 Yiheyuan Road, Beijing, 100871, China;

18 7School of Forestry, Northeastern Forestry University, Harbin, 26 Hexing Road,
19 150040, China.

20
21 ***Corresponding authors:**

22 J. William Munger (jwmunger@seas.harvard.edu)

23 XiuLing Man (mannefu@163.com)

24
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39

40 **Abstract**

41 High latitude boreal forests are experiencing dramatic changes in climate and
42 hydrology. It is not clear how boreal forests will adapt to hydrological change or how
43 stable they will be to extreme climate fluctuations and shifts in ecosystem water
44 availability (EWA; residuals between precipitation and evapotranspiration). Although
45 there have been numerous studies in North American and European boreal forests, the
46 Siberian boreal region is underrepresented. Moreover, Siberia is dominated by
47 deciduous conifers (larch) that may have different response to shifting hydrology than
48 boreal evergreens do. We observed evapotranspiration (ET) by eddy covariance
49 technique and transpiration (T) by sap-flow probes on a subsample of trees within the
50 flux-tower footprint through two growing seasons in a larch forest in northernmost
51 China. Ecosystems at the margins of their zone could be amongst the first to
52 experience significant shifts in structure and function. At this site there have already
53 been signs of permafrost degradation and more frequent temperature and precipitation
54 anomalies. The canopy-dominant larch accounted for half the total T fluxes. The
55 remaining 50% was distributed evenly among intermediate and suppressed trees. T is
56 the dominant subcomponent in ET , where overall T/ET varies of 66%–84% depending
57 on precipitation patterns. In dormant and early growing seasons, T still constitutes a
58 majority of ET even though the canopy foliage is not fully developed because cold
59 soil creates a negative soil to air vapor pressure gradient that impedes evaporation.
60 However, in the peak growing season, excess precipitation reduces T while providing
61 sufficient wetness for surface evaporation. ET from standard data product based on
62 MODIS satellite reflectance underestimates tower ET by 17%–29%. Solar-induced
63 chlorophyll fluorescence measured by satellite is well correlated with tower ET ($r_2 =$
64 0.69–0.73) and could provide a better basis for regional ET extrapolations. A global
65 comparison of data for 2000–2018 period reveals that boreal forests not only have the
66 smallest annual MODIS ET but also the least EWA compared to temperate and
67 tropical forests. Also, even though boreal deciduous and evergreens have comparable
68 annual ET , their T/ET and EWA are distinct. This work highlights how short-term
69 precipitation extremes may shift ecosystem function and structure by changing EWA
70 through exported runoff. Sites along boreal ecotones are critical to observe for signs
71 of shifts in their structure, function, and response to climate anomalies.

72

73 **Keywords** Ecosystem stability; Precipitation extreme; Climate anomaly; Phenology;
74 Boreal forests

75

76 **1 Introduction**

77 **Overview on the significance of studying boreal water exchanges**

78 Evapotranspiration (*ET*) is a significant component in global hydrological cycling
79 and land-atmosphere energy balances (Good et al., 2015). *ET* includes the exchange
80 of water from the land surface to the atmosphere by the abiotic evaporation and biotic
81 transpiration (Katul et al., 2012). *ET* is critical but challenging to accurately predict
82 because it links two different pathways of water vaporization and also interacts with
83 the climate closely (Shukla et al., 1990). Forest ecosystems are the essential engine in
84 the terrestrial water exchanges (Jasechko et al., 2013; Schlaepfer et al., 2014;
85 Schlesinger and Jasechko, 2014). Thus, they are of great interest to foresters,
86 ecologists, hydrologists, and climatologists for understanding *ET* and transpiration
87 across a range of spatial scales and various ecosystem types (Kool et al., 2014).
88 Quantifying the ratio of transpiration to *ET* (T/ET) contributes to understanding
89 ecosystem carbon-water coupling and cycling (Austin et al., 2004). Accurate
90 representation of *ET* partitioning and its influence on land-atmosphere patterns is
91 essential for realistic climate simulations (Lawrence et al., 2007).

92 Increasing studies have explored the T/ET across many terrestrial ecosystems by
93 field experiments (Yepez et al., 2003; Moran et al., 2009; Scanlon and Kustas, 2010;
94 Cavanaugh et al., 2011; Tian et al., 2011; Raz-Yaseef et al., 2012; Sun et al., 2014;
95 Kool et al., 2016), meta-analysis (Jasechko et al., 2013; Coenders-Gerrits et al., 2014;
96 Schlesinger and Jasechko, 2014; Wang et al., 2014; Wei et al., 2017), and ensemble
97 modeling (Lawrence et al., 2007). Based on a global synthesis (Schlesinger and
98 Jasechko, 2014), transpiration accounts for 61% of terrestrial *ET*. Boreal forests,
99 temperate forests and tropical rainforests make distinct contributions (reanalyzed in
100 Figure S1).

101 **Climate change influences the stability of boreal ecosystems**

102 Over the past 30 years climate change in northern ecosystems, has led to a decline
103 of ground albedo following more frequent wildfires and permafrost degradation
104 (Tchebakova et al., 2009). The release of previously frozen carbon by permafrost
105 thawing may amplify climate change (Melillo et al., 2002; Schuur et al., 2015). On
106 annual to decadal time scales, climate change strongly alters precipitation and
107 temperature patterns (Lotsch et al., 2003). Regional temperature-precipitation
108 complexity influences vegetation phenology (Goulden et al., 1996; White et al., 1999;
109 Nolan et al., 2018), in particular, boreal forest's mortality and hydraulic vulnerability
110 are very sensitive to climate extremes (Peng et al., 2011; Way et al., 2013). The
111 Siberian ecoregion is one of the largest boreal forest in the world (Watson et al.,
112 2018), which serves as a significant carbon sink. Hence, amplified climate change,
113 degraded permafrost, and the increased chance of temperature-precipitation anomalies
114 are challenging the resilience and stability of boreal forests in Siberia (Tchebakova et
115 al., 2009). Warmer temperatures have contributed to expanding boreal forest into
116 tundra at the northern tree-line (Esper and Schweingruber, 2011), and simultaneously
117 to the expansion of temperate forest into boreal forest at the southern margin (Evans
118 and Brown, 2017).

119 Northeast China, which is at the southern margin of the Siberian zone, contains 3.8
120 $\times 10^5$ km² of permafrost that is experiencing severe degradation (Jin et al., 2007). The
121 degradation is occurring in the discontinuous permafrost zone as well as continuous
122 permafrost zones. Permafrost degradation influences ecosystem energy and mass
123 exchange that will impact overlying vegetation (Yi et al., 2014). Much of this region
124 is covered by vulnerable larch forests intermingled with early succession birch/aspens
125 stands and scattered Mongolian Scots pines.

126 **Water flux measurements and remotely-sensed indicators**

127 Many direct and indirect measurement approaches have been developed for
128 estimating *ET* and its sub-components since the 1970s (Kool et al., 2014), including
129 micro-lysimeters (Deguchi et al., 2008), sap-flow measurements (Granier et al.,
130 1985), isotopic tracing (Jasechko et al., 2013), eddy covariance (Wilson et al., 2001),
131 hydrological process-based modeling (Estevez et al., 2009), energy balance residual
132 (Amiro, 2009), and carbon-water correlation (Scanlon et al., 2010). Direct
133 measurements by the eddy covariance method are considered to be the most reliable
134 and accurate approach so far (Tian et al., 2011). Indirect methods usually depend on
135 several theoretical assumptions that add additional uncertainties (Shi et al., 2008).

136 Algorithms based on remote-sensing data provide estimates of *ET* on regional to
137 global scales (Velpuri et al., 2013). For instance, the Moderate Resolution Imaging
138 Spectroradiometer (MODIS) *ET* is a mature data product with good spatial resolution
139 and extended temporal coverage and has been used for global synthesis already
140 (Schlesinger and Jasechko, 2014). However, it has not been extensively evaluated
141 against field observations in the widespread, critical Siberian ecosystem (Mu et al.,
142 2011). Observations are needed to test whether empirical ecosystem parameters based
143 on evergreen conifers can be applied to the boreal deciduous conifers in Siberia
144 (Reich et al., 1998).

145 Meanwhile, solar-induced chlorophyll fluorescence (SIF) is an emerging technique
146 that observes vegetation's photosynthetic processes by quantifying the fluorescent
147 radiation emission (Joiner et al., 2014). Increasing studies show SIF has potentials to
148 track ecosystem water and carbon dynamics from the single site to the globe (Lee et
149 al., 2013; Guan et al., 2015; Sun et al., 2015; Lu et al., 2018). However, there are no
150 studies available to determine whether SIF is a reliable indicator for *ET* estimations in
151 boreal larch forest.

152 **Uncertainties on boreal water budget evaluations**

153 Currently, there are two knowledge gaps in global *ET* and transpiration evaluations.
154 First, uncertain *ET* budgets in boreal larch ecozones. Larch is the deciduous conifer
155 that dominates Siberian boreal forests with a total area of 6.8×10^6 km² (Abaimov,
156 2010; Pan et al., 2011). Ecosystem functional traits of deciduous larch are distinct
157 from evergreen conifers (Reich et al., 1998). However, global ecohydrological studies
158 have not included larch at all, which has left uncertainties in evaluating regional to
159 global water budgets (Jasechko et al., 2013; Schlesinger and Jasechko, 2014).

160 Second, daily transpiration fluxes estimated from sap-flow measurements remain
161 uncertain because the relative contributions of xylem refilling and actual water losses
162 to nocturnal sap flow (Fisher et al., 2007) have not been adequately determined for

163 many plant functional types, in particular, boreal larch forest.

164 Furthermore, the response of boreal larch's hydrology to microclimate driving
165 forces (e.g., temperature, humidity, radiation, etc.) is still unclear. In general, forest *ET*
166 and transpiration are controlled strictly by atmospheric vapor pressure deficit and
167 solar radiation (Motzer et al., 2005; Mackay et al., 2007). Temperature is an essential
168 regulator for some ecosystem types (Law et al., 2002), though its tight correlation
169 with radiation, especially at high latitudes, makes it difficult to separate these
170 influences. Shifting of inter-annual precipitation patterns will also introduce
171 uncertainties into correlations between forest hydrology and climatic drivers
172 (Trenberth, 2011; Endo et al., 2017).

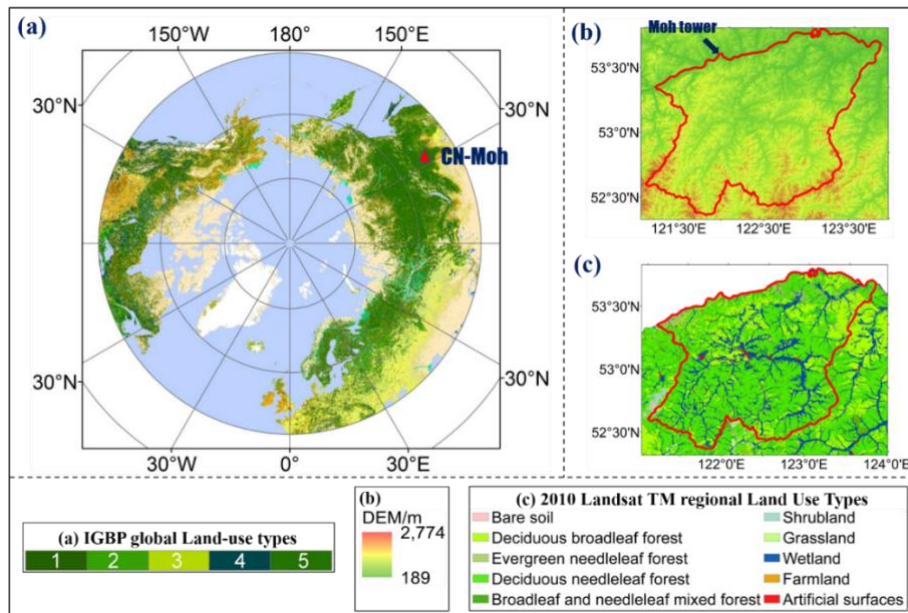
173 In this study, we focus on the overarching objective of understanding how *ET* in
174 boreal larch forest responds to precipitation changes. Secondly, we seek to distinguish
175 xylem refilling from actual water loss so that nighttime sap-flow observations can be
176 accurately accounted. Additionally, we examine key methodological issues that affect
177 the ability to estimate water budgets over larger spatial scales and longer intervals.
178 How well can *ET* for boreal larch forest be estimated from remote-sensing data such
179 as MODIS reflectances or satellite derived SIF? How well can machine learning
180 algorithms estimate *ET* and enable prediction at places and times where measurements
181 fail?

182 **2 Methodology**

183 **2.1 Site description**

184 The Mohe (Moh) ecological research station (53°27'59" N, 122°20'06" E) is located
185 in the Mohe County, Northeast China (location in the globe and IGBP types were
186 highlighted in Figure 1a). The study area is within the zone of Eurasian continuous
187 permafrost. The site near the Mohe tower is fairly flat with the elevation of 290 m a.s.l
188 in the Amur River valley (Figure 1b), and it also lies at the northern part of the
189 Greater Hinggan Mountains and near the southern edge of Siberia with similar flora
190 and climate. Regional mean annual temperature (1959–2017) is -5.5 °C and the mean
191 annual precipitation in the same period is 460.8 mm with approximately 70%
192 occurring as rainfall during warmer months (i.e., June to September). Also, Shannon
193 Diversity Indices of the satellite pixel in Figure 1c showed the site is uniform
194 (Richness = 2.0, Evenness = 0.24).

195 The native forest is dominated by Dahurian larch (*Larix gmelinii*), along with the
196 evergreen Scots pine (*Pinus sylvestris* var. *mongolica*), and two deciduous broadleaf
197 trees, white birch (*Betula platyphylla*) and aspen (*Populus davidiana*). Understory
198 vegetation includes ledum (*Ledum palustre*), lingonberry (*Vaccinium vitis-Idaea*),
199 saussurea (*Saussurea japonica*), Dahurian rhododendron (*Rhododendron dauricum*),
200 and blueberry (*Vaccinium uliginosum*). Tree heights of larch were 12.4 ± 4.2 m (mean
201 \pm s.d.) with 11.1 ± 5.9 cm diameter at breast height and 7.1 ± 5.2 m² canopy area in
202 2015 (Table S1). Understory heights were 0.1–2.0 m with 3%–76% ground coverages.
203 Land use/land cover change in Mohe was insignificant from 1990–2010 (Figure S2).
204 In 2010, larch forest covered 56% lands in Mohe county (1.0×10^4 of 1.8×10^4 km²)
205 with negligible change (expansion by ~ 100 km²) since 1990.



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Figure 1. Panels (a): Landcover map for the boreal-arctic region north of 45° (Globcover2009: http://due.esrin.esa.int/page_globcover.php). **Panel (b):** Digital-Elevation-Model (DEM) of Mohe. **Panel (c):** Regional land use map in Mohe county based on Landsat TM images in 2010. The red polygon is the border of Mohe county. The 30-m spatial resolution DEM data was downloaded from the Geospatial Data Cloud (<http://www.gscloud.cn/>). The 30-m spatial resolution land use map was self-interpreted from the Landsat 5 TM images (<https://landsat.usgs.gov/>) and plotted on the base map. IGBP legend in (a): (1) Closed to open (ground cover > 15%) broadleaf evergreen and/or semi-deciduous forest (tree height > 5 m), (2) Closed (> 40%) broadleaf deciduous forest (> 5 m), (3) Open (15–40%) broadleaf deciduous forest (> 5 m), (4) Closed (> 40%) needleleaf evergreen forest (> 5 m), (5) Open (15–40%) needleleaf deciduous or evergreen forest (> 5 m).

2.2 Sap-flow measurements and estimations

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The field campaign was from April 15 to September 25 in 2015 (Julian day of 105–268) and 2016 (Julian day of 106–269). All trees with > 5 cm diameter at breast height in the plot were counted within a 400 m² sampling plot (good representative of entire tower footprint) for sap-flow measurements situated 100 m north of the Mohe tower. There were 94 larch and 1 pine, and no birch or aspen in the sampled plot. Sap-flow measurements for pines and birches were ignored due to their low abundance. Intraspecies competition (dominant, intermediate, and suppressed) of larch was determined by degrees of canopy dominance (higher tree height with larger canopy area).

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A 20 cm length increment borer (Haglöf, Långsele, Sweden) was used to drill growth cores (bark to pith; in a south to north direction) from 20 randomly selected larches. Sampled cores were dried at 70 °C for 48 h and polished by 240 and 800 sandpaper. LINTABTM linear tree-ring measuring stage (RINNTECH, Heidelberg, Germany) was used for estimating sapwood widths and forest age (~51 yr) from processed cores. We found that the sapwood area (tissues for water transportation) was tightly correlated with the diameter at breast height, that was fit to a linear law

236 (Figure S3d). We selected nine uncored larches (out of 34) for the sap-flow
237 measurement by thermal dissipation probes (details in Supplementary Materials
238 Section 3), and each group of dominant, intermediate, and suppressed trees had three
239 samples. Total magnitudes of transpiration in each class of the plot were upscaled by
240 total sapwood area and sap-flow density (*Eq 1* and *Eq 2*).

$$241 \quad T_{\delta} = J_{s_{-\delta}} \times \frac{A_{s_{-\delta}}}{A_G}, \delta = d, i, s \quad \text{Eq 1}$$

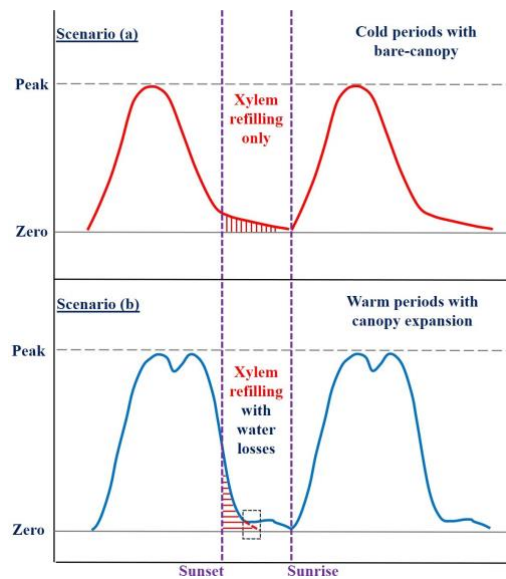
$$242 \quad T_{tot} = T_d + T_i + T_s \quad \text{Eq 2}$$

243 where: T_{δ} is the magnitude of transpiration (mm h⁻¹) in tree classes of dominant (*d*),
244 intermediate (*i*), and suppressed (*s*); $J_{s_{-\delta}}$ is sap-flow density of each tree class (m³ m⁻²
245 d⁻¹); $A_{s_{-\delta}}$ is the total area of sapwood in each tree class; A_G is the plot area (m²); T_{tot} is
246 the sum of transpiration from all larch classes.

247 **2.3 Nighttime sap-flow determinations**

248 As indicated in Figure 2, nighttime sap flow was based on the intensity of
249 environmental driving forces over different phenophases (Fisher et al., 2007).
250 Assumed scenarios including: **(a)**. When nighttime vapor pressure deficit (VPD, the
251 primary driver for transpiration), was near zero, as generally occurred during the cold
252 periods, we assume that any observed sap-flow during night is refilling only. The diel
253 pattern of sap-flow was marked by a rise soon after sunrise, declining with decreasing
254 solar radiation in the evening and tailing off to zero within a couple of hours after
255 sunset (Fisher et al., 2007). **(b)**. When nighttime VPD was elevated, as generally
256 during warm periods, both transpiration and xylem refilling are expected to occur.

257 Note that Fisher et al. (2007) considered two approaches to analyze the nighttime
258 partitioning: (1) Projected refilling into the transpiration period (yields an upper
259 bound for refilling), and (2) Back-extrapolated transpiration into the refilling period (a
260 lower limit on refilling, upper limit on water loss). Theoretically, refilling water losses
261 should occur together during the transition and the separation will be biased by either
262 abovementioned approach (outlined by dark-grey box in Figure 2 Scenario b). We
263 apply a compromise approach that distinguishes transpiration from xylem refilling
264 based on the time of an inflection point in sap-flow curve rather than extrapolate the
265 curves for each process during the transition. The xylem refilling is subtracted from
266 nighttime sap-flow measurements for analysis and quantification throughout this
267 study unless mentioned specifically.
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270 **Figure 2.** Schematic graphs indicate the scenarios on patterns and determinations of
 271 nighttime sap-flows. Red crosshatching indicates the portions of xylem refilling. In
 272 scenarios (b), dotted line rectangle highlighted the tail of xylem refilling is under the
 273 water loss (overlapped two processes; Fisher et al., 2007).

274 **2.4 Net water flux measurements**

275 An integrated three-dimensional sonic anemometer and open-path infrared gas
 276 analyzer, the IRGASON (Campbell Scientific, Logan, USA) installed at 36 m (24 m
 277 above the mean canopy top) at the top of the Mohe tower measured wind velocity,
 278 sonic temperature, and water vapor concentrations at 10 Hz. The high-frequency data
 279 were stored on CR3000 datalogger (Campbell Scientific, Logan, USA). Raw flux data
 280 were processed by the EddyPro (Li-Cor, Lincoln, USA), and a post-processing *R*
 281 package “*FREddyPro*” (<https://www.cran.r-project.org/packages/REddyProc>) was
 282 applied to verify the reliability of the output data and remove outliers. Energy closure
 283 analysis is discussed in the supplementary materials (Figure S4), and the correction of
 284 latent heat fluxes followed Mauder et al. (2007; 2013) with the Tovi software (Li-Cor,
 285 Lincoln, USA).

286 *ET* and other fluxes including momentum, sensible heat, and CO₂ were calculated
 287 from the high-frequency data over 30-minute intervals. Results for carbon flux
 288 dynamics are presented in a concurrent study focused on Siberian larch’s carbon
 289 dynamics (Liu et al., 2019). Axis rotation for tilt correction used the method of planar
 290 fit with no velocity bias (van Dijk, 2004). Turbulent fluctuations were calculated by
 291 applying the block-averaging method.

292 Observations under low turbulence situations, which have unreliable eddy flux
 293 values, were also eliminated by the *u** threshold (Moving Point Test; Gu et al., 2005).
 294 The *u** threshold (0.13 m s⁻¹ at Mohe site) was derived from the entire day but not
 295 daytime/nighttime only. Approximately 42% and 49% of the total observed fluxes had
 296 acceptable *u** values in 2015 and 2016, respectively. Also, it shows that averaged
 297 water flux in low turbulence situations is 0.78 ± 0.23 mmol m⁻² s⁻¹, while in high
 298 turbulence situations is 1.23 ± 0.31 mmol m⁻² s⁻¹.

299 The final flux QA/QC step included manual checks of daily and monthly variations
300 and monthly diurnal-nocturnal distributions of *ET*. Water vapor vertical profiles were
301 not measured, so the storage term for water flux could not be calculated. We estimated
302 a lower limit for the canopy storage by assuming a constant water vapor mixing ratio
303 beneath the IRGASON for deriving the ecosystem *ET* (Papale et al., 2006). Because
304 the forest canopy and trunk space were fairly open, this assumption was reasonable.

305 We found the manufacturer's firmware for correcting the spectroscopic effect of the
306 IRGASON's gas analyzer used a slow-response sensor to capture the air temperature,
307 which did not provide enough temporal resolution to accurately correct CO₂
308 concentrations when air temperature was below -1.0 °C (Wang et al., 2016). In place
309 of the standard firmware, we calculated a correction based on the fast-response
310 temperature sensor (sonic anemometer) applied to the individual 10 Hz CO₂ values.

311 The IRGASON is designed to perform well in high-humidity conditions using
312 innovative features, including: (1) Gas analyzer windows are polished, slanted at an
313 angle, and coated with a hydrophobic material to prevent water from collecting on
314 their surfaces; (2) Wicks are used on the windows to promote capillary action and
315 move water away from the window edges; (3) Heaters in the inlet are turned on to
316 help minimize data loss because of precipitation and condensation events.

317 **2.5 Microclimate measurements**

318 Meteorological measurements were made at 35 m height above the ground. Air
319 temperature and relative humidity were measured using passively-shielded HMP155
320 probes (Vaisala, Vantaa, Finland). Wind speed was measured using a 010C wind speed
321 sensor (MetOne, Grants Pass, USA). Photosynthetically-active radiation and total net
322 radiation were measured at 23 m height by a LI190SB (Li-Cor, Lincoln, USA) and an
323 NR01 sensor (Hukseflux, Delft, Netherlands), respectively. Four CS650 integrated
324 soil temperature, moisture and conductivity probes (Campbell Scientific, Logan,
325 USA) were placed into the ground at 5, 10, 20, and 40 cm depths to measure the
326 profile soil temperatures and volumetric soil water contents. Microclimatic
327 measurements were sampled every 5 min and saved as 30 min averages by the
328 CR3000 datalogger (Campbell Scientific, Logan, USA).

329 Site level liquid precipitation was measured at 23 m height of the tower using a
330 TE525 unheated tipping bucket rain gauge (Campbell Scientific, Logan, USA).
331 However, this single point measurement was likely an underestimate because the solid
332 precipitation (e.g., snowfall, hail, and sleet) was not collected, and wind-blown
333 precipitation was inefficiently sampled (Figure S5). Unless noted otherwise we used
334 the regional precipitation data from the 0.5° by 0.5° national precipitation
335 compilations (Data Center of China Meteorological Administration; CMA) to
336 represent the regional condition of precipitation. CMA precipitation was calculated
337 based on ground meteorological stations and upscaled to a 40-km spatial resolution.
338 Also, in this study, rainy days were defined as their daily cumulative precipitation \geq
339 1.0 mm, and the rest were defined as non-rainy days.

340 **2.6 Full list of adopted remote sensing data**

341 In addition to the remote sensed products for land use types and site level geo-
342 information illustrated in Figure 1 (i.e., GlobCover, DEM, and Landsat TM), we also

343 adopted another four satellite products and two publicly released long-term weather
344 datasets, including MODIS leaf area index (LAI; MCD15A3H; Myneni et al., 2015),
345 MODIS fractional snow cover (FSC; MOD10A1; Hall and Riggs, 2016), Global
346 Ozone Monitoring Experiment-2 (GOME-2) SIF (Joiner et al., 2014), MODIS *ET*
347 (MOD16A2; Mu et al., 2007), Global Precipitation Climatology Project (GPCP)
348 monthly precipitation dataset (Adler et al., 2003), and Berkeley Earth monthly
349 averaged/bias-corrected air temperature (Rohde et al., 2003). Moreover, MODIS
350 products showed in this study were all pre-processed by The Oak Ridge National
351 Laboratory Distributed Active Archive Center (<https://daac.ornl.gov/>). GOME-2 SIF
352 in the study was normalized by the solar zenith angles (method described by Luus et
353 al., 2017).

354 Moreover, phenophase in boreal larch forest is determined by the curvature-change
355 model using the MODIS LAI (Zhang et al., 2003). Specifically, transition dates
356 correspond to the time at which the curvature values exhibit local minima or maxima.
357 During greenness, the two maximum values correspond to the onset of greenness
358 increase (leaf-onset) and the onset of greenness maximum (leaf full-expansion).
359 Similarly, during periods of LAI decrease the two minimum values identify the onset
360 of greenness decrease (leaf-senescence) and the onset of greenness minimum (leaf-
361 dormant). The confidence limits (s.d.) of the season lengths and transition days was
362 estimated by using Monte-Carlo simulations. Based on fitted parameters of curvature-
363 change model, we ran 500-times permutations of these random variables to calculate
364 all possible outcomes of the transition dates. Note that s.d. is considered if the
365 difference between length of greenness periods across 2015 and 2016 is significant or
366 still within the uncertainty and time resolution of the MODIS LAI derived transition
367 dates.

368 **2.7 Machine learning prediction of water fluxes**

369 In order to integrate flux observations to the longer time intervals needed for
370 evaluating water and carbon budgets, gaps introduced by instrument downtime and
371 invalid measurements must be filled in. Machine learning has been well developed for
372 predicting *ET* since the early-2000s (Whitley et al., 2009; Mehdizadeh, 2018). In this
373 study, we adopted RandomForest for the gap-filling (Breiman, 2001). RandomForest
374 is originally designed for purposes of ecological classifications (Cutler et al., 2007),
375 while its full application to the ecological regressions evolved afterward by others
376 (Iverson et al., 2007; Labrière et al., 2016; Schwalm et al., 2017). RandomForest
377 constructs multiple trees (i.e., information assembled clusters) individually, and each
378 tree grows with a randomized subset of predictor variables and further yields
379 predictions independently. Afterward, all decisions will be aggregated and then
380 averaged to produce the final predictions. More importantly, RandomForest possesses
381 good data economy and performances, especially in dealing with the less
382 computationally intensive tasks or smaller datasets (e.g., usually 35% gaps in flux
383 measurements; Falge et al., 2001).

384 RandomForest calculations were performed in *R* (*RandomForest* package;
385 <https://www.cran.r-project.org/web/packages/randomForest/>). Motivated by the
386 parameters that represent energy exchange, atmospheric turbulence, and ambient

387 water supply in the Penman-Monteith equation (Sun et al., 2014; Sun et al., 2016), we
 388 selected the following fundamental variables (e.g., directly measured) to train the
 389 algorithm: (1) Net radiation, air temperature, and soil temperature, which reflect
 390 energy distributions and changes; (2) Wind speed and relative humidity, which reflect
 391 the atmospheric turbulence above the canopy; (3) Soil water content, which reflects
 392 the available water resource from the ground level. Only the fundamental variable is
 393 accepted, because the function-based variables (e.g., vapor pressure deficit versus
 394 temperature and humidity) will affect machine learning's decision on speculating their
 395 functional relationships and result in the overfitting (Schaffer, 1993).

396 Observed *ET* in 2015 and 2016 ($n = 13\ 408$; excluded 4038 gaps) was randomly
 397 divided into non-duplicated training dataset ($n = 7500$) and evaluation dataset ($n =$
 398 3500). The training dataset was input for RandomForest. Evaluation dataset was kept
 399 for verifying the functionality of RandomForest predictions. Settings and summary
 400 statistics for the RandomForest simulation are given in Table 1. The following
 401 evaluations concluded that RandomForest was an ideal solution for the predictions of
 402 *ET* in boreal larch forest (adjusted $r^2 = 0.97 \pm 0.01$; Figure 3a). Though the peak-
 403 values of *ET* were slightly underestimated by the algorithm, the overall seasonal
 404 patterns were still well simulated (Figure 3b).

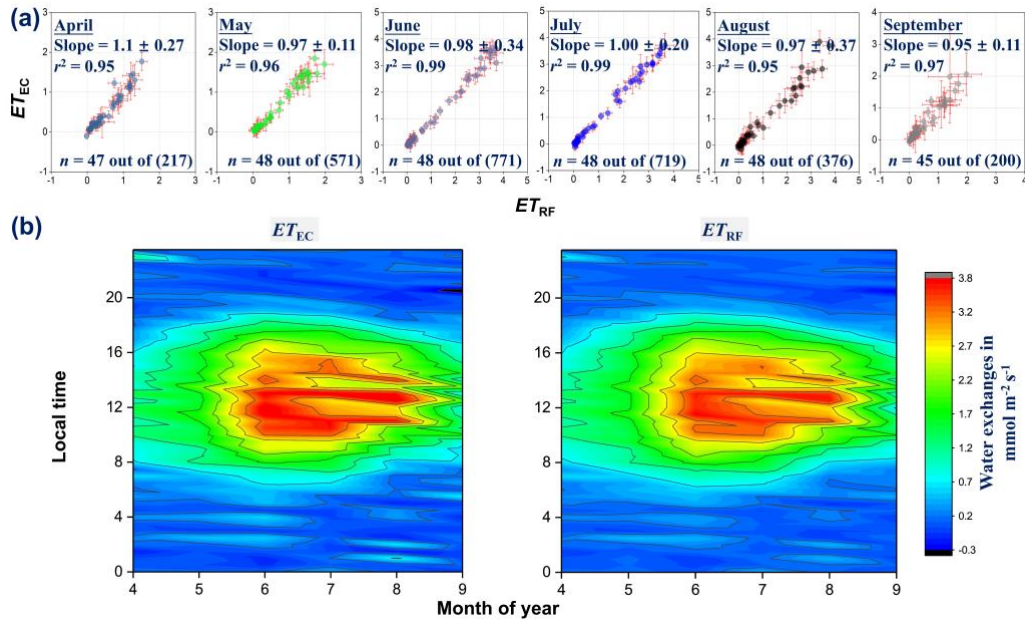
405

406 **Table 1.** Statistics on the eigenvalues of RandomForest simulation. Note that the
 407 statistics on the bottom of table were based on a 7500 sized training dataset.
 408 Abbreviations of *n*tree and *m*try represent the number of trees and the number of
 409 variables per level, respectively. Unit of root mean square error (RMSE) is in mmol
 410 $\text{m}^{-2} \text{s}^{-1}$. OOB is the out of bag error rate, representing the proportion of data that are
 411 not used for the individual regression tree-model building. Mean decrease in Accuracy
 412 (MDA) and mean decrease in Gini index (MGI) are two importance scores output
 413 from the RandomForest algorithm, in short, the larger the value, the more significant
 414 the result is. Abbreviations of T_s , SWC, R_{net} , *WS*, T_a , and *RH* are soil temperature at 5
 415 cm depth, weighted volumetric water content (from 5 and 40 cm depths), upper
 416 canopy total net radiation, upper canopy wind speed, upper canopy air temperature,
 417 and upper canopy relative humidity, respectively.

Preferences and statistics	No. of training data	No. of gap-filled	No. of ntree	No. of mtry	RMSE	OOB
	7500	4038	1000	2	0.62	0.44
Importance scores	T_s	SWC	R_{net}	<i>WS</i>	T_a	<i>RH</i>
MDA	0.48	0.28	2.21	0.21	0.47	0.27
MGI	1 379.7	1 005.7	7 415.1	1 160.5	2 340.8	1 221.7

418

419



420

421 **Figure 3.** Evaluations on the RandomForest predicted ET (ET_{RF}) by the tower
 422 observation (ET_{EC}) with a 2500 sized dataset. Data in panels (a) and (b) are
 423 aggregated between 2015 and 2016. In panel a: samples n are binned from a data
 424 clusters (size showed in parentheses) by the time of day with the format of median \pm
 425 s.e. Also the uncertainties (s.e.) of linear slopes are showed in each regression plots.
 426 Different filling colors indicate data periods (e.g. the month).

427 3 Results

428 3.1 Microclimates in China's boreal larch forest

429 Generally, the climate in the southern edge of Siberia is influenced by the semi-
 430 permanent continental Siberian Anticyclone, which keeps the region cool and dry. The
 431 continuous permafrost at Mohe site thaws to a depth of 50–100 cm (shallower on
 432 shady slopes and deeper on sunny slopes) by the end of summer, which is within the
 433 range (50–110 cm) for zonal permafrost thaw depths monitored at a boreal site near
 434 the southern Greater Hinggan Mountain (50°N) and a Central Siberian boreal site (56°
 435 N; Figure S6).

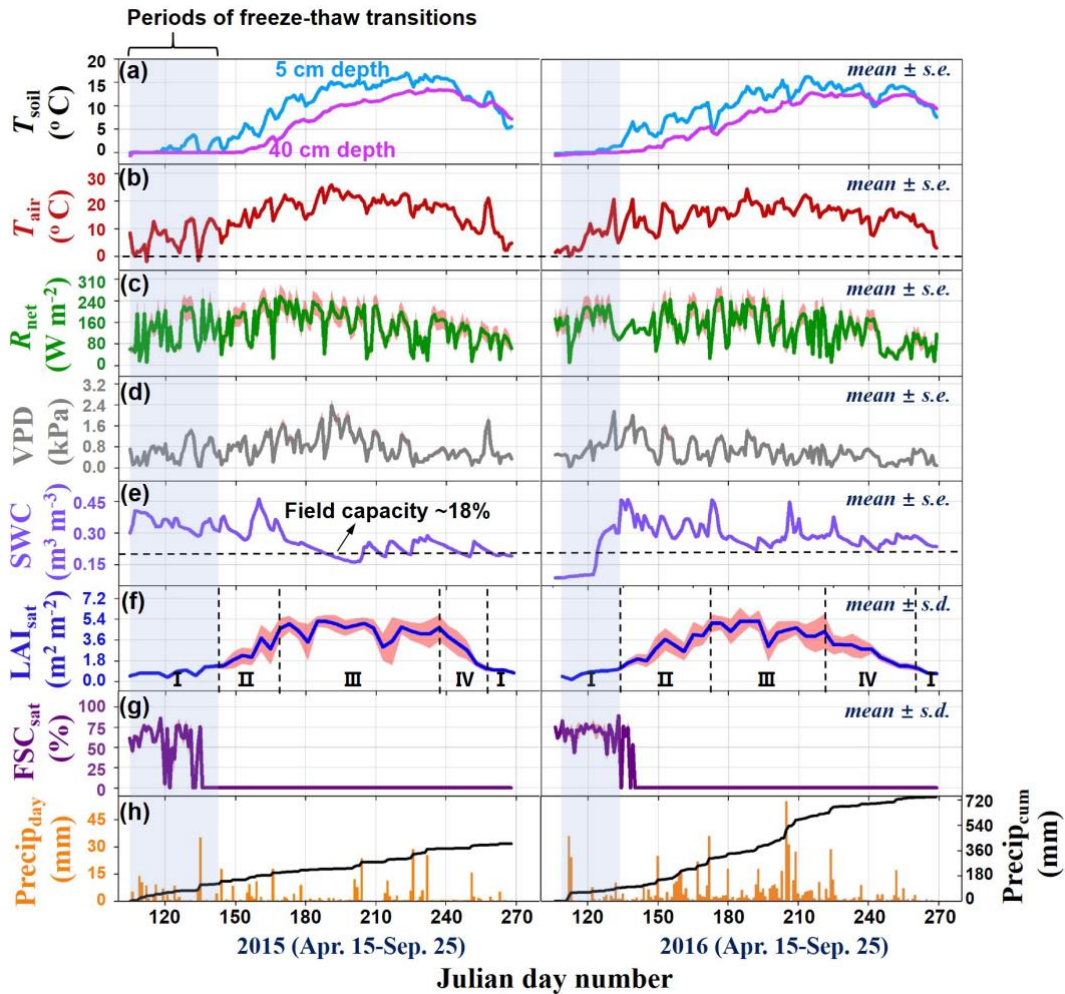
436 A total of 328 days field observations (from April 15 to September 25 with 164
 437 days of each year) were included in this study (Figure 4). No observations were made
 438 for the remaining periods when solar power was insufficient, and site access was
 439 challenging due to the harsh weather. As highlighted in Figure 5h, differences in
 440 precipitation patterns and total accumulation for the observed periods were very
 441 obvious between 2015 (409.9 mm) and 2016 (743.5 mm). Precipitation in 2015 was
 442 dominated by events yielding $< 1 \text{ mm d}^{-1}$, while precipitation intensity in 2016 was
 443 shifted towards larger events (Figure S5). Total annual precipitation of 510.2 mm in
 444 2015 was similar to the regional mean precipitation of 460.8 mm (1959–2017), while
 445 2016 with 792.1 mm was an anomalously wet year. Also, we observed the 2016
 446 anomaly is among the top three precipitation extremes in the entire data records.

447 Precipitation links closely to patterns of air temperature and VPD (Figure S7). The
 448 heavier precipitation at Mohe in 2016 was part of a larger scale precipitation anomaly

449 that was identified by GPCP (Figure 5). So that, consistency of GPCP and CMA
450 precipitation anomaly in 2016 highlighted this precipitation excess is a regional
451 feature and not just local to Mohe (Figure 5a, Figure S5). Precipitation and
452 temperature anomalies were more frequent since 2002. The precipitation anomaly was
453 slightly lagged behind the temperature anomaly (Figure 5a, c). Taken together,
454 precipitation patterns in Mohe site, in particular during the growing season, were
455 closely related to the local temperature (Figure 5e), as the warm air masses carry
456 moisture and synoptic weather patterns affect them together.

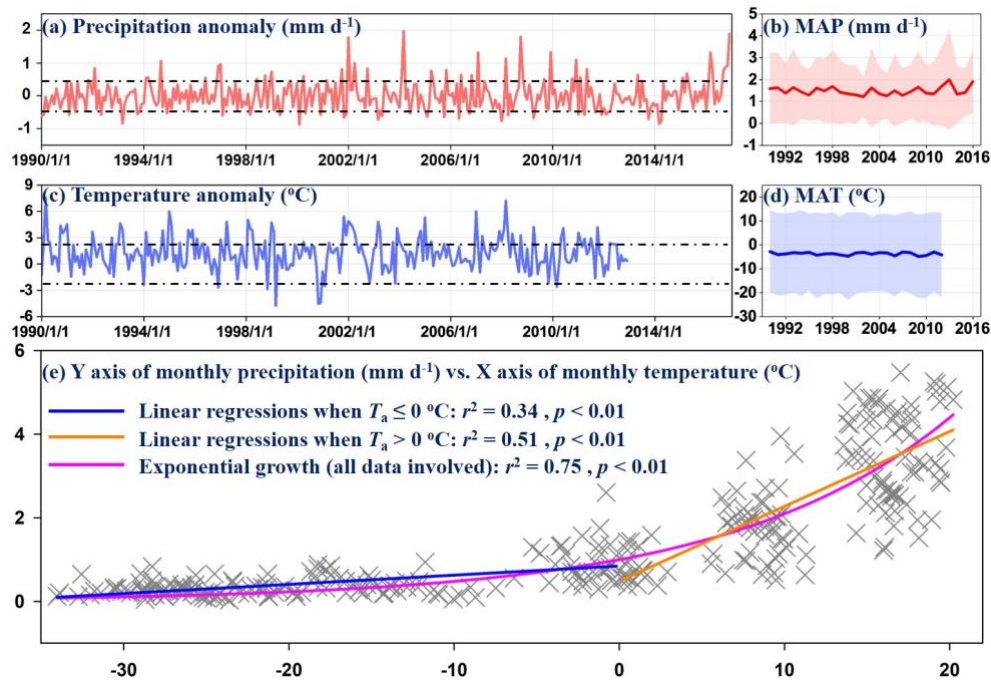
457 Each year was divided into four canopy-phenophases of leaf-dormancy, leaf-onset,
458 full leaf expansion, and leaf-senescence based on the MODIS LAI (Figure 4f). The
459 growing season length (from leaf-onset to leaf-senescence) in 2015 was 111 (± 3.3 ;
460 s.d.) days and was extended to 120 (± 4.7) days in 2016, where this extension was due
461 to longer leaf-onset and leaf-senescence phenophase instead of the shorter leaf-full-
462 expansion compared to 2015. The main meteorological variables were significantly
463 different between the phenological stages (Table S2).

464 2015 starts with wet soil but dries over summer, and enters dormant season still dry
465 until thaw and new precipitation saturate the soil in 2016 spring (Figure 4e). Soil
466 moisture in 2016 recharged quickly from its dry condition to near saturation following
467 heavy rain on the day of year 112 that fell on the melting snowpack (Figure 4g).
468 Except for the dry period in dormant phenophase of 2016 soil moisture in the
469 remainder of both years was at or above the field capacity.



470

471 **Figure 4.** Time series of microclimatic variables during two-years field campaigns.
 472 T_{soil} , T_{air} , R_{net} , VPD, SWC, LAI, FSC, $\text{Precip}_{\text{day}}$, and $\text{Precip}_{\text{cum}}$ represent for soil
 473 temperature, air temperature, total net radiation, vapor pressure deficit, weighted soil
 474 water content (5 and 40 cm depths), satellite leaf area index, satellite fractional snow
 475 cover, daily precipitation, and cumulative precipitation, respectively. I, II, III, and IV
 476 with the dashed line in panel (f) identify the four canopy phenophases, i.e., leaf-
 477 dormant, leaf-onset, leaf full-expansion, and leaf-senescence, respectively. The leaf
 478 dormant period is additionally identified by gray shading. The black dashed line in
 479 panel (e) indicates field capacity in boreal larch forest measured during growing
 480 season of 2015. MODIS LAI and FSC were both retrieved with 2.5 km by 2.5 km
 481 pixels by centered on the Mohe flux tower. FSC was smoothed using a *R loess* filters.



482

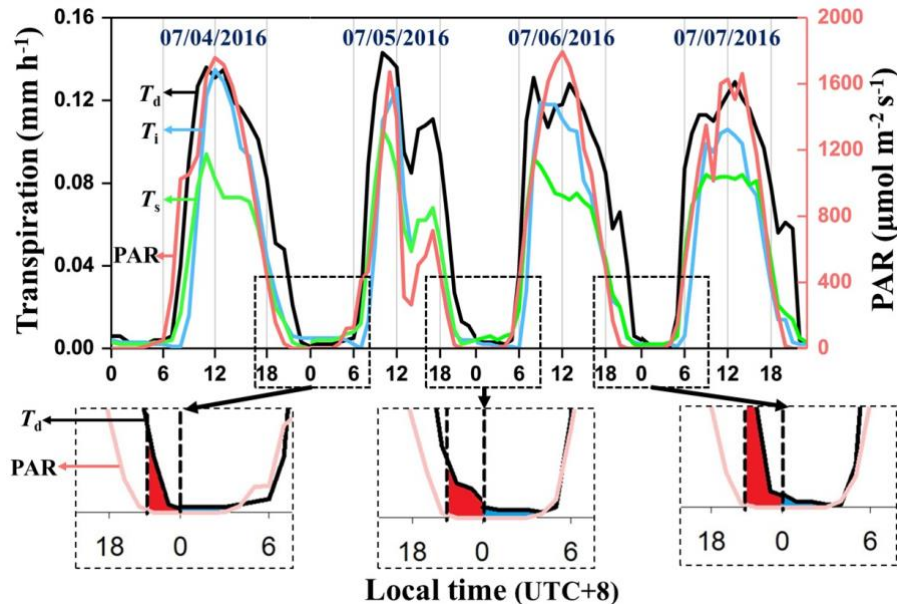
483 **Figure 5. Panels (a)–(b):** Monthly precipitation anomalies and annual mean
 484 precipitation rates given by GPCP precipitation for the 2.5° by 2.5° grid surrounding
 485 Mohe (Adler et al., 2003) are shown for 1990–2016. **Panels (c)–(d):** Monthly
 486 temperature anomalies and mean annual temperature rates from the Mohe station
 487 given in Berkeley Earth air temperature dataset are shown for 1990–2012. **Panel (e):**
 488 Monthly precipitation against monthly temperature for periods of 1990–2012.
 489 Anomalies are computed by calculate the long-term mean from the 1990–2016 data,
 490 and then subtracting the overall mean from the data for a given month and year.
 491 Throughout the text, we consider abnormal climate events to be those years in which
 492 temperature or precipitation anomalies exceed 0.5 s.d. (black dashed lines) of the
 493 interannual temperature or precipitation variabilities (Diffenbaugh et al., 2015).

494 3.2 Nighttime sap-flow separation of boreal larch forest

495 We observed sap-flow fluctuations at the end of dormant phenophase (day of year
 496 142 in 2015 and 133 in 2016) prior to the leaf-out suggesting that sap transport
 497 supports metabolism in unopened buds (Figure S8). Nighttime sap-flows primarily
 498 responded to nighttime above-canopy VPD. The influence of nighttime above-canopy
 499 air temperature was minimal (Figure S9). Also, we found that dominant larch was
 500 more sensitive to environmental conditions, as measured above the canopy, than the
 501 intermediate and suppressed larch that are in the subcanopy.

502 The separation of the nighttime sap-flow was run for the entire dataset and then
 503 visualized by taking four-day data sequences as examples (Figure 6). Boreal larch
 504 forest usually needs four hours to refill the xylem water deficit after the intense
 505 daytime water losses during non-dormant phenophase. By our definitions (described
 506 in Section 2.3), nighttime sap-flows can be numerically distributed into xylem
 507 refilling and real water losses. So that the larger proportion of xylem refilling, the
 508 smaller percentage of water losses. We quantified the xylem refilling in dominant
 509 larch groups, which constituted 95% of integrated nighttime sap-flows (i.e., 5% real

510 water losses). While for intermediate and suppressed larches, xylem refilling
 511 accounted only for 80% (20% water losses). These 5%–20% nighttime water losses
 512 equaled to approximately 2%–9% of total daily transpiration. The smaller proportion
 513 of xylem refilling in intermediate and suppressed larches indicated that being in the
 514 shade protected these trees from drying by direct radiation and may have reduced the
 515 VPD they were exposed to.



516
 517 **Figure 6.** Snapshot of nighttime sap-flow separation during the growing season.
 518 Patterns of sap-flow were primarily controlled by changes in vapor pressure deficit
 519 (Figure S9). The set of images across the lower portion of the panel focus on the
 520 nocturnal periods identified by shading. Xylem refilling and water losses are shown
 521 by the red and blue shading, respectively. Determination approach explained in
 522 Section 2.3.

523 3.3 Precipitation and phenology regulate water fluxes

524 Variations of precipitation, phenophase and their interactions influenced water
 525 fluxes in boreal larch forest (Table 1). We applied principal component analysis to
 526 identify the dominant variables (out of 11 possible measurements) that would be
 527 considered in the microclimatic regulation analysis (Table S3). The first principal
 528 component accounted for 40.4% of the total variance. The variables that correlated the
 529 most with the first principal component were air temperature (0.62), vapor pressure
 530 deficit (0.53), surface soil temperature (0.45), and net radiation (0.45).

531 Different levels of annual precipitation input could modify the effect of
 532 microclimate on water fluxes (Figure S10). For instance, *ET* was significantly
 533 controlled by soil temperature in both 2015 (normal year) and 2016 (the wetter year).
 534 While transpiration was only influenced by soil temperature in the normal year. On
 535 the other hand, the evaporation which includes contributions from canopy
 536 interception, forest floor, exposed soil, and any surface water, calculated as difference
 537 between *ET* and transpiration, was related to soil temperature in the wetter year but
 538 not in the normal year.

539 The principal component results were corroborated by stepwise regressions (Table
540 2). The stepwise regression further highlighted that transpiration of three larch canopy
541 classes responded differently to microclimate. Excess precipitation introduced
542 uncertainty to the prediction of transpiration (e.g., overall lower r^2 in the wetter year).
543 Also, we found transpiration would increase with VPD (positive signs) because of
544 increased atmospheric demand.

545 In general, analysis revealed that, air temperature, net radiation, and VPD are
546 primary controllers of biotic hydrological processes (e.g., transpiration) in this boreal
547 larch ecosystem. It is a common recognition that forest *ET* and transpiration are
548 controlled strictly by air temperature, net radiation, and VPD (Law et al., 2002;
549 Motzer et al., 2005; Mackay et al., 2007). Furthermore, we saw controls by VPD were
550 even stronger in the year with precipitation extremes, while the influence of air
551 temperature and net radiation declined when water availability was in excess. Shifting
552 of inter-annual precipitation patterns will introduce uncertainties into correlations
553 between forest hydrology and climatic drivers (Trenberth, 2011; Endo et al., 2017).
554

555 **Table 1.** An analysis of variation procedure with tower and repeated sap-flow
556 measurements and the Tukey's Honestly Significant Difference test here to analyze
557 the fixed effects of phenophase, precipitation, observational years, and their
558 interactions on water flux components. Formula of glm() function in R: 'WFC ~
559 phenophase + precipitation + year', where WFC is water flux component listed below.
560 T_d , T_i , T_s , T_{tot} , and ET represent transpiration from dominant trees, intermediate trees,
561 suppressed trees, ecosystem transpiration fluxes, and ecosystem evapotranspiration,
562 respectively. Df , Adj SS, Adj MS are degrees of freedom, the adjusted sum of squares,
563 and the adjusted mean sum of squares, respectively. Statistical terms (in particular the
564 interactions) that were not significant in the final model do not appear in the table.

Component	Effect	df	Adj SS	Adj MS	F	P
T_d	Phenophase	3	10.571	3.524	13.99	0.000
	Precipitation	1	2.664	2.664	10.57	0.001
	Year	1	3.486	3.486	13.84	0.000
	Pheno \times Year	3	1.957	0.652	2.59	0.053
	Precip \times Year	1	1.321	1.321	5.24	0.023
	Residuals	317	79.854	0.252		
T_i	Phenophase	3	6.481	2.160	26.57	0.000
	Precipitation	1	1.124	1.123	13.82	0.000
	Year	1	0.178	0.178	1.69	0.194
	Precip \times Year	1	0.333	0.333	4.09	0.044
	Residuals	320	26.022	0.081		
	T_s	Phenophase	3	5.015	1.672	29.47
Precipitation		1	0.536	0.536	9.44	0.002
Year		1	0.314	0.314	5.54	0.019
Pheno \times Year		3	0.776	0.259	4.56	0.004
Precip \times Year		1	0.314	0.314	5.53	0.019
Residuals		317	17.985	0.057		
T_{tot}	Phenophase	3	73.609	24.536	24.86	0.000
	Precipitation	1	14.332	14.332	14.52	0.000
	Year	1	4.497	4.4974	4.56	0.034
	Precip \times Year	1	4.535	4.5355	4.60	0.033
	Residuals	320	315.775	0.987		
	ET	Phenophase	3	225.361	75.120	75.48
Precipitation		1	17.580	17.576	10.40	0.001
Year		1	3.070	3.070	3.08	0.030
Pheno \times Year		3	10.201	3.400	3.42	0.018
Residuals		318	316.501	0.995		

565

566 **Table 2.** Stepwise regression analysis of water flux with microclimatic variables
567 across 2015 and 2016. Microclimatic variables were selected by the PCA analysis
568 (Table S3). σ indicates uncertainties of each coefficient. NA represents no
569 corresponding parameter involved.

Year	Regression equation	σT_{soil}	σT_{air}	σR_{net}	σVPD	r^2	F -value	p -value
2015	$T_d = 0.12 + 0.02 T_{soil} + 0.00 R_{net} + 0.49 VPD$	0.03	NA	0.01	0.17	0.85	294.2	0.00
	$T_i = -0.17 + 0.01 T_{soil} + 0.02 T_{air} + 0.00 R_{net} + 0.09 VPD$	0.02	0.01	0.00	0.01	0.92	452.6	0.00
	$T_s = -0.06 + 0.01 T_{air} + 0.00 R_{net} + 0.09 VPD$	NA	0.07	0.05	0.12	0.89	413.2	0.00
	$E = 0.49 - 0.04 T_{soil} + 0.09 T_{air} - 1.35 VPD$	0.07	0.11	NA	0.28	0.20	13.22	0.00
	$ET = 0.28 + 0.11 T_{air} + 0.01 R_{net} - 0.72 VPD$	NA	0.08	0.00	0.57	0.62	87.5	0.00
2016	$T_d = -0.11 + 0.02 T_{air} + 0.00 R_{net} + 0.32 VPD$	NA	0.01	0.06	0.21	0.67	103.3	0.00
	$T_i = -0.29 + 0.02 T_{soil} + 0.00 R_{net} + 0.32 VPD$	0.03	NA	0.00	0.18	0.81	215.6	0.00
	$T_s = -0.13 + 0.01 T_{air} + 0.00 R_{net} + 0.34 VPD$	NA	0.07	0.12	0.27	0.74	145.9	0.00
	$E = 0.17 + 0.11 T_{soil} + 0.01 R_{net} - 1.16 VPD$	0.08	NA	0.09	0.78	0.47	45.9	0.00
	$ET = -0.34 + 0.16 T_{soil} + 0.02 R_{net}$	0.12	NA	0.00	NA	0.66	152.9	0.00

570

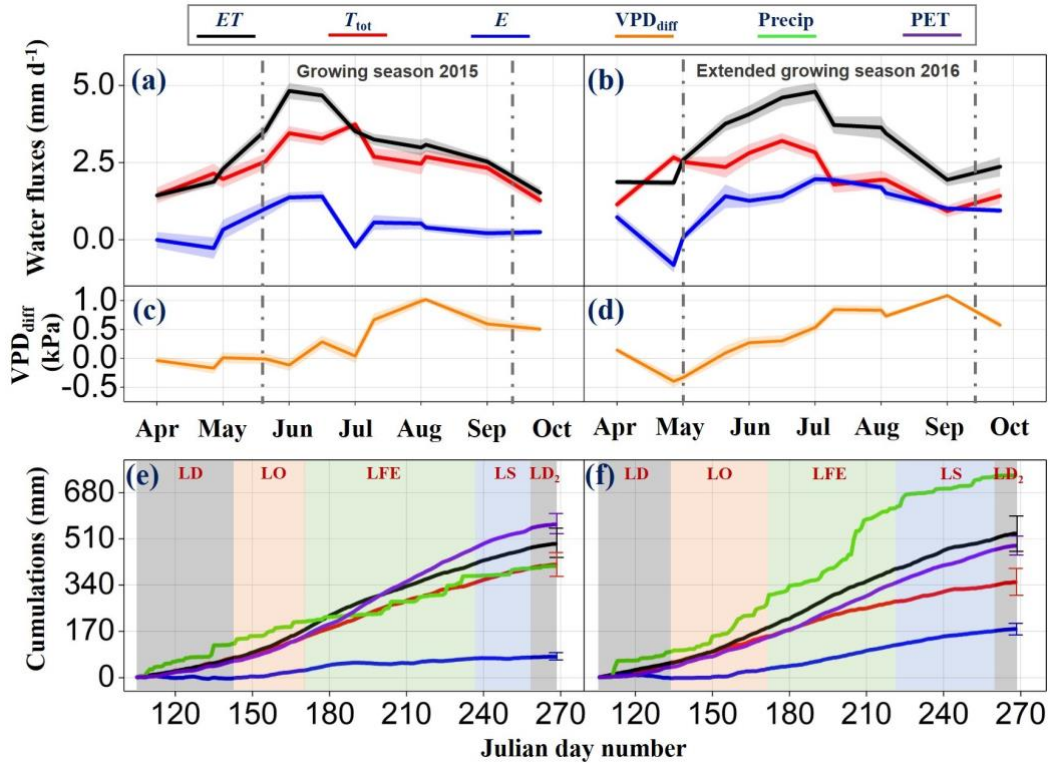
571 3.4 Patterns and partitioning of water fluxes

572 As seen in Figure 7a–d, the Mohe site forest exchanged 491.6 ± 133.4 (95% CI)
573 mm of water by *ET* in 2015, including 415.1 ± 82.1 mm as transpiration (where $T_d =$
574 221.7 mm, $T_i = 96.6$ mm, and $T_s = 96.9$ mm) and 76.47 ± 51.3 mm as evaporation. In
575 2016, *ET*, transpiration, and evaporation were 529.4 ± 146.9 mm, 350.5 ± 101.3 mm
576 (in which $T_d = 178.7$ mm, $T_i = 88.4$ mm, and $T_s = 83.4$ mm), and 178.9 ± 45.6 mm,
577 respectively. Evaporation was influenced by the soil to atmosphere vapor pressure
578 gradient. The positive soil to air vapor pressure gradient promotes water evaporation,
579 while negative gradients (particularly in the spring when the air warms faster than the
580 soil (see Figure 4 a-b) impedes evaporation or even cause condensation at the soil
581 surface (dew). For example, we occasionally observed the negative evaporation of
582 2015 (indicating water moves from atmosphere towards the ecosystem; Table S4)
583 during the leaf-dormant and leaf-onset periods, which is the result of the negative
584 gradient (Figure 7c).

585 Despite the considerable increase in precipitation for 2016, *ET* for 2016 was only
586 marginally greater than in 2015 (Figure 7e-f; $F_{1, 325} = 2.96$, $p = 0.09$). Evidently,
587 increased precipitation is not promoting a higher *ET*. In both years, *ET* sums were
588 mostly comparable to potential evapotranspiration sums (2015: $F_{1, 325} = 510.83$, $p =$
589 0.04 ; 2016: $F_{1, 325} = 601.89$, $p = 0.06$), indicating *ET* is energy limited but not water
590 limited. Together, these results suggested that the ecosystem stored enough water to
591 support all *ET* demands that could occur and rarely if at all was *ET* suppressed by lack
592 of available moisture. Again, note that totals of water fluxes in this study are not from
593 the full year, so the water deficits that occur are only in periods from April to October
594 (when we had field observations). But the year-round water deficits can be balanced
595 by winter snowpack. On average winter snowpack is up to 30% of the annual
596 precipitation (< 138 mm). Also, field observations during some dry anomalous years
597 are needed in the future to test if water availability is ever limiting at this site.

598 Figure 8a shows that transpiration by dominant larch accounted for at least 50% of
599 total ecosystem transpiration, while the remainder was almost equally split between
600 intermediate and suppressed larches. To examine the direct influence of precipitation
601 we binned the data into rainy and non-rainy days. *ET* magnitude increased on the
602 rainy days (Figure 8b–c). But transpiration had a distinctly lower contribution on
603 rainy days. Either increased evaporation of intercepted precipitation in the canopy or
604 suppression of transpiration by reduced VPD, temperature, and solar radiation
605 associated with rainy weather (or both together) account for the shift in *ET*
606 partitioning.

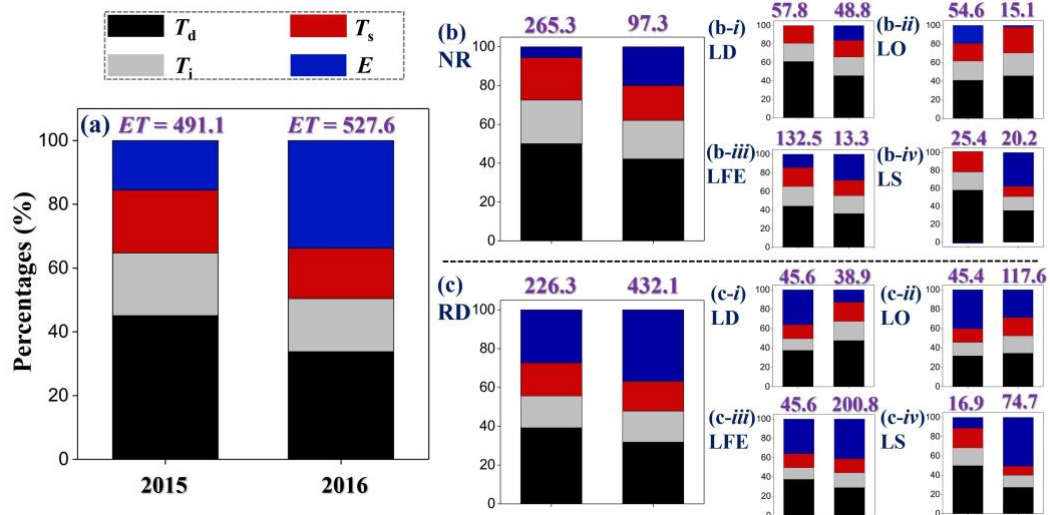
607 *ET* partitioning is also affected by phenology. During the full leaf expansion
608 phenophase when LAI was at its peak and active photosynthesis requires maximum
609 stomatal opening, transpiration accounted for a larger fraction of *ET*. During the early
610 growing season before larch needles were fully expanded, but foliage was present in
611 understory shrubs and scattered evergreen conifers, transpiration contributed a large
612 fraction to *ET* on non-rainy days (Figure 8b series). Even though there is abundant
613 standing water at the surface and sunlight penetrates to the ground, evaporation is
614 suppressed in this season because the temperature of those wet surfaces is maintained
615 below the dewpoint by soil frost below (see Figure 4 a–b, Figure 7 c–d).



616

617 **Figure 7. Panel (a)–(d):** time-series variations of ET , total transpiration (T_{tot}), and
618 and evaporation (E) and vapor pressure gradient between soil and air (VPD_{diff}) in 2015
619 and 2016. **Panel (e)–(f):** Cumulative water flux components precipitation and
620 potential evapotranspiration (PET) through 2015 and 2016 observation period.
621 Evaporation (E) derived from difference between ET and T_{tot} , where T_{tot} comprises
622 transpiration from dominant (T_d), intermediate (T_i), and suppressed (T_s) trees. Lines
623 connect the mean values \pm s.e. of the biweekly bin (panel a–d), and cumulative sums
624 were derived from daily values (panel e and f). 95% CI was given at the end of each
625 cumulative by colored +/- segments. Growing season marked with vertical grey
626 dashed lines throughout panel (a)–(d). Leaf-dormant (LD), leaf-onset (LO), leaf full-
627 expansion (LFE), and leaf-senescence (LS) identified with shaded color backgrounds
628 in panel (e) and (f). PET (purple lines) is the amount of evaporation and transpiration
629 that would occur if a sufficient water source were available (estimated by the
630 Thornthwaite equation in R package 'SPEI'; <http://sac.csic.es/spei>). Snapshots of
631 interannual dynamics of daily observed water fluxes and VPD_{soil} are shown in Figure
632 S11.

633



634

635 **Figure 8.** *ET* partitioning among the tree-size classes and evaporation in 2015 (left
636 columns) and 2016 (right columns) for all data (panel a) and binned according to
637 presence/absence of rain (daily rain ≥ 1 mm) (panel b and c), and by phenophase (b-*i*,
638 b-*iv* and c-*i*, c-*iv*). NR and RD are non-rainy days and rainy days respectively. LD,
639 LO, LFE, and LS are phenophase of leaf-dormant (including both LD and LD₂
640 showed in Figure 7e), leaf-onset, leaf full expansion, and leaf-senescence. The overall
641 magnitude of *ET* (mm) for each data subset is given by the number above the
642 partitioning column.

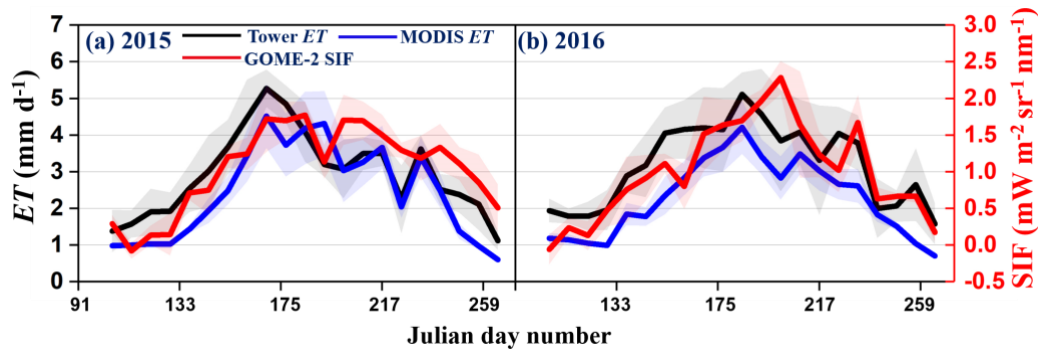
643 3.5 Estimates of water flux from remote sensing observations

644 To evaluate the influence of spatial variability, we aggregated the MODIS *ET*
645 values over different spatial scales from single pixel (0.5 km), through larger regions
646 covering 10 km, 20 km, and 40 km surrounding the tower location. Spatial
647 aggregation did not change the seasonal pattern appreciably (Figure S12a, b). Cloud
648 screening introduces frequent data gaps in the 500 m MODIS *ET* that was retrieved
649 from the single pixel surrounding the tower (e.g., Table S5). In particular note the
650 gaps during leaf-full-expansion period of 2016. Unless otherwise specified all
651 comparisons of *ET* observations and remote-sensing-based estimates will use the 40
652 km aggregation to smooth out variability and reduce the influence of cloud gaps.

653 In 2015, MODIS *ET* provided agreed with *ET* measured at the tower during the
654 growing season (Figure 9a). But in 2016, MODIS *ET* underestimated *ET* over the
655 entire growing season (Figure 9b). Pixel quality is negatively correlated with
656 precipitation (Table S6), and the percentage of MODIS pixels that passed QA/QC was
657 reduced in 2016 compared to 2015 (Figure S12d). We also observed decreased
658 accuracy of MODIS *ET* for each studied year especially during dormant seasons.
659 Snow reflectance and clouds are known issues that compromise the reflectance
660 measurements that MODIS *ET* is based on. When many pixels in are rejected for
661 cloud contamination, the remaining pixels may not be as representative of the tower;
662 in particular persistent weather patterns could result in the cloud-free pixels having
663 less precipitation than those that were usually cloudy.

664 As yet no algorithm has been developed to compute *ET* from SIF, but we plot tower
665 *ET* together with SIF in both 2015 and 2016 (Figure 9 a, b) to compare their seasonal

666 patterns. Tower *ET* and SIF had similar seasonal patterns, and tower *ET* had
 667 reasonable correlations (r_2) 0.69 and 0.73 with SIF in 2015 and 2016, respectively
 668 (Figure S13). Comparison of MODIS *ET* and the tower observations (Table S5) shows
 669 that on average MODIS *ET* is less than the tower *ET* by approximately 23% with low
 670 to 17% and up to 29% during the observation periods in 2015 and 2016, respectively.



671

672 **Figure 9. Panel (a)–(b):** Evaluation of correlation between MODIS *ET* or GOME-2
 673 SIF (mean \pm s.e.) and *ET* observed at the Mohe site. MODIS *ET* and GOME-2 SIF
 674 were both retrieved for 40 km blocks centered on the Mohe tower.

675 4 Discussion

676 4.1 Flux gap filling and nighttime transpiration in boreal larch forest

677 Although it is necessary for calculating flux integrals over intervals longer than
 678 individual half hours, gap filling of flux data remains challenging and introduces large
 679 uncertainty. Some frequent-used conventional methodologies (e.g., mean diurnal
 680 variations and lookup tables, etc.), account for flux dependence on a limited number
 681 of factors and capture the mean value (Falge et al., 2001). However, ecosystems are
 682 complex systems that comprise multiple biotic and abiotic components that may
 683 interact with each other through both linear and non-linear ways (Levin, 1998). In
 684 recent decades, machine learning has become the focus of much attention, and is a
 685 state-of-the-art solution for extracting patterns and insights from earth system data
 686 (Reichstein et al., 2019), due to the wide range of its applicability and the ease with
 687 which it can handle both linear and non-linear sectors of the data simultaneously (Lek
 688 and Guégan, 1999; Olden et al., 2008).

689 We show here preliminary applications of machine learning in the gap-filling of
 690 water fluxes for boreal forests. The machine learning approach is chosen because it
 691 does not impose any functional forms (Lek and Guégan, 1999; Olden et al., 2008).
 692 Trained by a set of fundamental micrometeorological variables, the RandomForest
 693 algorithm yields overall accurate predictions of *ET* in boreal larch forest. However,
 694 we emphasize that its application to other ecosystems still needs to be further verified
 695 and spatial and temporal scales of input data are an important factor limiting the
 696 performance of machine learning algorithms (Whitley et al., 2009; Mehdizadeh,
 697 2018).

698 Furthermore, this is the first study that determines the nighttime xylem refilling in
 699 boreal larch forest (80%–95% depending on tree classes). Nighttime xylem refilling is
 700 a critical plant functional trait that enables them to transpire faster than water can be

701 transported from the roots for short periods during the day. During the early morning,
702 sap-flow that initiates in the upper trunk utilizes stored water, and the onset of sap-
703 flows at the base of the trunk is usually delayed by minutes to hours (Scholz et al.,
704 2008). Using stored water allows plants to keep stomata open and continue to
705 photosynthesize longer. Also, releasing stored water buffers the negative water
706 potential in xylem, while without buffering, the high tension could result in cavitation,
707 which causes unrepairable damage to the water transport system. Hydraulic resistance
708 in the stem can limit water transport from roots to foliage (Whitehead, 1998). The
709 situation where evaporation demand exceeds water uptake will lead to a hysteresis
710 response between soil water supply and stoma water losses (Buckley, 2005). The
711 depleted water stores are thus to be refilled during the night.

712 **4.2 Microclimates in China's boreal larch forest**

713 Microclimatic conditions are significant factors affecting structure and function of
714 forest ecosystems (Arx et al., 2013), whereas feedbacks of the forest ecosystem to
715 climate systems will modify precipitation regimes (Lotsch et al., 2003). Patterns of
716 precipitation, length of various phenophase and their interactions limit levels of
717 transpiration and *ET* of larch. We saw air temperature, net radiation, and vapor
718 pressure deficit are the main external variables that control water exchange processes
719 in boreal larch forest (also consistent with Law et al., 2002; Motzer et al., 2005;
720 Mackay et al., 2007). The microclimatic regulations on *ET* during precipitation
721 extremes are more complicated than normal situations because *ET* is the sum of
722 transpiration and evaporation that are independently influenced by microclimate with
723 varied relative contributions (Katul et al., 2012). Our results suggest precipitation
724 excess will alter ecosystem energy limits and change evaporation and transpiration
725 partitioning in boreal larch ecosystem (Zeppel et al., 2008).

726 Transpiration itself will be reduced in situations of precipitation excess
727 (Wullschleger and Hanson, 2006). Evaporation of boreal larch will be increased at the
728 same time because it was calculated as the residuals between *ET* and *T*. Meanwhile,
729 wet surface is often associated with the periods of excess precipitation, so there is
730 adequate water to be evaporated. In this study, precipitation came in the form of
731 frequent cloudy days with small amounts of precipitation for each event (Figure S5).
732 But receiving excess precipitation by way of infrequent but larger events might have a
733 different effect on the ecosystem. Also, data showed evaporation is controlled
734 primarily by the soil-air vapor pressure gradient (Wehr et al., 2017), but not direct
735 effects from radiations, temperatures, and humidity. Altogether, *ET* comes close to the
736 ideal PET value and does not go up particularly for the excess precipitation year. This
737 represented that the ecosystem water transportation is likely energy limited but not
738 moisture limited. However, observation during a year with below regional
739 precipitation is needed to test some assumptions related to precipitation deficiency.

740 High latitude ecosystems are prone to ecological degradation due to the recent
741 frequent and intensified climate anomalies (Melillo et al., 2002; Schuur et al., 2015).
742 At Mohe site, we observe temperature-precipitation anomalies are more frequent in
743 the most recent decade. Arctic-amplification, the more rapid warming of high northern
744 latitudes compared to the global average (Cohen et al., 2014), could be a controlling

745 factor. Also, the long-term local temperature-precipitation record illustrates a tight
746 correlation between temperature and precipitation patterns, especially during larch's
747 growing season. The coupling between precipitation and temperature introduces
748 complexity in ecosystem response to climate change (Nolan et al., 2018).

749 **4.3 Transpiration and evapotranspiration in response to the exterior and interior** 750 **forcing**

751 Vegetation phenology, as the long-term adaption and acclimation strategy to local
752 climate, is another factor affecting larch's water fluxes. Patterns and partitioning of
753 water flux components in larch's leaf full-expansion phenophase can determine the
754 intra-annual water budgets due to a larger amount of exchanged water. While in non-
755 growing season or the earlier growing season, trees have low transpiration capacity
756 because they have no leaves, or their stomatal are not yet functional. The precipitation
757 baselines and phenology patterns in 2015 and 2016 represent contrasting conditions
758 within the range of interannual variability. Transition from dormancy to spring onset
759 in 2016 has higher air temperature and soil water content than in 2015, which provide
760 insights on how the boreal larch forest would respond in a warmer and wetter climate
761 that has an earlier spring onset and extended growing season length (GSL; Menzel
762 and Fabian, 1999; Chen et al., 2005).

763 There is a recent trend of increasing GSL in northern ecosystems (White et al.,
764 1999). Small changes in the timing of spring growth will strongly control the annual
765 carbon exchange (Goulden et al., 1996), and due to coupling of carbon and water
766 fluxes in terrestrial ecosystems, annual water budgets should respond similarly. For
767 instance, we show the extended GSL in 2016 is possibly the result of excess moisture
768 (i.e., precipitation anomaly), where 1 day increase in GSL will change larch's *ET* by
769 1.2% (2015)–1.4% (2016). This variation is considerably larger than the 0.2%
770 alteration for eastern US deciduous broadleaf forest reported by White et al., 1999.

771 In addition, intraspecies competition (to light) influences *ET* partitioning. Overall,
772 transpiration, as the most robust engine for water transportation in boreal larch forest,
773 is consistent with other ecosystems worldwide (Jasechko et al., 2013; Schlesinger and
774 Jasechko, 2014). When separating the total transpiration flux into three larch classes,
775 we find dominant trees are the major transpiration source in the ecosystem *ET*, while
776 intermediate and suppressed trees account for the reminder with comparable levels.
777 Transpiration from leaf-full-expansion phenophase is governing the *ET* partitioning,
778 while excess precipitation increases the proportion of evaporation in the total *ET*.

779 Precipitation utilization by ecosystems, also recognized as ecosystem hydraulic
780 strategy, is a complex function influencing the ecosystem functional traits (West et al.,
781 2008). In this study, we define Ecosystem Water Availability (EWA) as the difference
782 between precipitation and *ET* (EWA_d for site-level evaluations in Figure 7) or as ratio
783 of *ET* to precipitation (EWA_r for uniform global comparisons). EWA contributes to
784 groundwater recharge and streamflow. Excess precipitation could maintain the
785 saturation in permafrost active layer, which affects whether the organic soils are oxic
786 or anoxic. EWA also affects active layer depth because water conducts heat more
787 effectively than air-filled pore spaces.

788 In boreal ecosystems which are accustomed to using all the water it gets through
789 the year would be strongly affected if permafrost melted and water could drain away
790 quickly. The water that pools in spring and additional water that becomes available
791 over summer as thaw progresses down to base of active layer would no longer be
792 available to support *ET* in late summer. At Mohe site, we observe *ET* slightly exceeds
793 precipitation in 2015 (−81 mm water deficit), whereas precipitation greatly exceeds
794 *ET* in 2016 (+215 mm water surplus). EWA also linked closely to terrestrial carbon
795 sequestrations. Boreal soils tend to be rich in organic matter (boreal evergreen vs.
796 Mohe: 92.6 vs. 107.6 Mg ha⁻¹; Data by GSOCmap on
797 <http://54.229.242.119/GSOCmap/> and by field experiments at Mohe) that decomposes
798 slowly on account of anoxic conditions when it is saturated together with freezing
799 conditions. Moreover, anoxia promotes methanogenesis while oxic conditions allow
800 methanotroph (Oswald et al., 2016). When soil drying allows aeration, decomposition
801 rates increase (Deluca and Boisvenue, 2012). For example, carbon accumulation in
802 the soil will be the greatest under water-saturated conditions (positive water deficit).
803 Soil drying (water deficit) and rewetting (water surplus) cycles enhance substrate
804 decomposition similar to freeze-thaw effects (Fierer et al., 2002).

805 **4.4 Estimation of water in boreal larch zones from remote-sensing data**

806 MODIS *ET* has been well evaluated and optimized by site-level measurements in
807 most vegetation ecosystems, with the notable exception of boreal larch forest (Mu et
808 al., 2007; Mu et al., 2011). The empirical ecosystem parameters based on evergreen
809 conifers that were applied to the boreal deciduous forests in Siberia would be
810 questionable (Reich et al., 1998). Previous studies using FLUXNET data showed that
811 MODIS *ET* generally correlated well with the observed *ET* in the conterminous US
812 but had residuals of 25%–50% (Velpuri et al., 2013). The relative error of MODIS *ET*
813 in a Finnish boreal forest was nearly 35% (Sánchez et al., 2007). We estimated, in
814 boreal larch forest, relative underestimation was approximately 17% for normal
815 precipitation situations but increased up to 29% for a year with excess precipitation.

816 Uncertainties in MODIS products can be attributed to the reduction in valid
817 observations and biased calculation algorithms. We find negative correlation between
818 MODIS quality and precipitation frequency, which suggests some errors might remain
819 after removing cloud-contaminated pixels and cloud screening is not completely
820 effective (Walther et al., 2016). The theoretical basis for the MODIS *ET* calculation
821 comes from the Penman-Monteith equation (Mu et al., 2007). P-M equation is a
822 commonly used hydrological model and predicts the rate of surface water fluxes by
823 following a single-layer ‘big-leaf’ approach (Estevez et al., 2009). It is a good tool in
824 some well-studied ecosystems with verified input parameters (Bond-Lamberty et al.,
825 2011; Tian et al., 2011; Sun et al., 2014; Sun et al., 2016). However, it may be biased
826 for data-poor ecosystems, for instance, the boreal larch forest.

827 Furthermore, spatial heterogeneity in the boreal larch landscape adds to uncertainty
828 when comparing MODIS *ET* to site-level *ET*. To mediate these uncertainties in boreal
829 larch forest, we adjust the MODIS *ET* magnitudes by errors we determined, and we
830 select the 40-km resolution pixels to reflect the regional *ET* situations for global
831 comparisons across different ecosystems. Our results also show that GOME-2 SIF is a

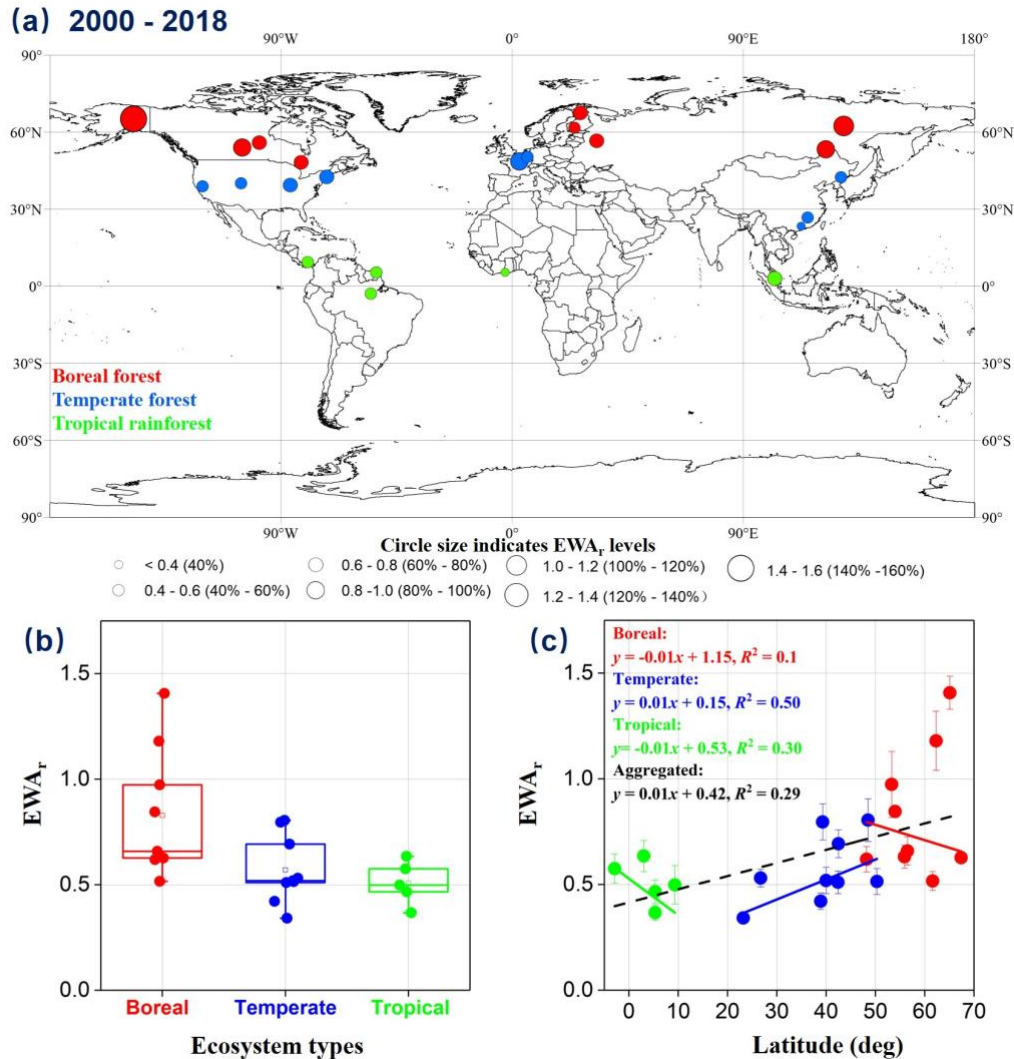
832 more reliable indicator that correlates tightly with ecosystem water fluxes in boreal
833 larch zones than MODIS *ET*. The close correlation between SIF and *ET* is consistent
834 with the previous study by Lu et al. (2018). This finding again reveals a great
835 potential of SIF in investigating the regional water dynamics and budgets.

836 **4.5 Water exchanges by boreal larch forest: a global perspective**

837 The ratio, T/ET , indicates the extent of vegetation dominance in an ecosystem. We
838 find growing season T/ET in boreal larch at Mohe was approximately 75% ($CI_{95\%} =$
839 18%) with a high of 84% in the normal precipitation year and a low of 66% in the
840 excess precipitation year. T/ET of boreal larch is comparable to the global average
841 ($\sim 65\%$) for boreal forest reported by Schlesinger and Jasechko (2014). North
842 American boreal spruce forests had a higher T/ET ratio from 70–80%, and European
843 boreal evergreen forests had lower T/ET of $\sim 50\%$. Accounting for boreal larch forest
844 based on the Mohe results would not shift the overall boreal forest average
845 significantly. Schlesinger and Jasechko's (2014) overall conclusion that tropical
846 rainforests and temperate forests have higher T/ET than boreal forests remains valid
847 (Figure S1).

848 We applied the EWA_r (Section 4.3) to understand vegetation roles in maintaining
849 ecosystem water budgets through 23 global boreal-temperate-tropical forest sites from
850 2000 to 2018 (Figure 10a). In order to allow for different timing of growing seasons
851 between biomes and duration of wet and dry or warm and cold seasons we compared
852 EWA_r on annual time increments (Figure S14). Although the range of EWA_r has
853 overlap across the ecosystem types, the general trend is for boreal $>$ temperate $>$
854 tropical (Figure 10b). EWA_r is negatively correlated with latitude in boreal and
855 tropical zones but not the temperate zone (Figure 10c). Although tropical rainforest
856 has the highest average *ET*, it has the lowest EWA_r . The low values indicate water
857 supply from precipitation exceeds ecosystem requirements. *ET* in boreal forests tends
858 to consume a large fraction of available water input putting them at a closer risk for
859 periods of moisture limitation (Table S7). Also, EWA_r in RU-SkP and US-Prr are
860 larger than 1.0 (Figure 10b). Persistent ecological disturbances, including
861 waterlogging and wildfires, are possibly major reasons that affect ecosystem water
862 availability in these two sites (Iwata, et al., 2011; Ohta et al., 2014).

863 The comparison of EWA_r points out variation in water across the boreal biome. The
864 two deciduous conifer sites, RU-SkP and CN-Moh, have comparable *ET* (365 ± 84 .
865 mm) to the other 7 evergreen conifers (389 ± 70 mm). However, EWA_r in deciduous
866 (1.1 ± 0.1) is distinctly larger than in evergreens (0.8 ± 0.3). The deciduous habit of
867 boreal larch may be an adaptation to generally dryer and cooler climate that evergreen
868 boreal conifers experience. We emphasize that vegetation roles in regional water
869 budgets across the entire boreal zone remains uncertain on account of sparse data
870 coverage throughout large portions of the boreal zone and in particular limited
871 information on the differences in water exchange properties between deciduous and
872 evergreen conifers.



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Figure 10. Panel (a): Overviews of EWA_r (ecosystem water availability; here calculated as the ratio of ET to precipitation) derived from 23 boreal-temperate-tropical forest sites in periods of 2000–2018. Specifically, MODIS ET is derived from 2000–2018, whereas precipitation is the mean annual precipitation recorded by FLUXNET. **Panel (b):** Comparisons of EWA_r (19 year averaged) between sampled sites. **Panel (c):** Regressions between EWA_r (error bars indicated \pm s.d.) and latitudes for sampled sites. Larger EWA_r represents ecosystem consumes more of available water. MODIS ET in two boreal larch sites (CN-Moh and RU-SkP) was increased by 23% (discussed in sections 3.6 and 4.4). The 23% adjustment will increase EWA_r values in two sites but not affect the overall ranks across different biomes.

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Supplementary data given in Table S7.

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5 Conclusions

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At the Mohe site, long-term precipitation patterns are positively correlated with the regional temperature patterns, which reveals a complex temperature-precipitation interaction. Predicting vegetation response to climate change will require consideration of the interactions and not simply response to temperature or moisture acting alone. The interaction between precipitation and phenology influences not only

891 the seasonal and inter-annual patterns of evapotranspiration but also its partitioning
892 between transpiration and evaporation.

893 Short-term precipitation extremes may shift ecosystem function and structure by
894 changing the distribution between water available to the ecosystem and exported as
895 runoff. Evapotranspiration slightly exceeds precipitation from April to October in
896 2015 and precipitation exceeds evapotranspiration by a factor of 1.4 during the same
897 period in 2016. Ecosystem water availability will further influence groundwater
898 recharge, streamflow, and heat transfer to the permafrost. When excess water keeps
899 soils in the active layer saturated, anoxic conditions that suppress decomposition rates
900 prevail and contribute to accumulation of organic carbon soils. Water deficits that
901 allowed enhanced soil aeration could accelerate decomposition.

902 Nighttime xylem refilling accounts for 80%–95% of nighttime transpiration, which
903 is 2%–9% of the daily total transpiration. Inadequate separation of xylem refilling and
904 actual water loss overestimates the vegetation water budgets, particularly during the
905 warmer but shorter growing seasons in the boreal zone. Transpiration is the dominant
906 water flux (66–84%) in boreal larch forest. From dormant to the early growing
907 seasons, transpiration constitutes the majority of evapotranspiration due to the
908 negative soil-air vapor pressure gradient; soil surface and standing water remain
909 colder than the dewpoints in the air. In the peak growing season when the canopy is
910 fully expanded, excess precipitation reduces transpiration, but enhances evaporation.
911 The combination of higher humidity and cooler foliage reduces vapor pressure deficit,
912 and less water is lost for a given stomatal opening. At the same time, precipitation
913 leaves more intercepted water on the canopy and trunk that can evaporate.

914 The ratio of transpiration to evapotranspiration in boreal larch forest is within range
915 observed at other boreal sites and is smaller than the ratios observed for temperate and
916 tropical forests. Tropical forests exchange more evapotranspiration but because they
917 have even higher precipitation inputs the fraction of water consumed by the
918 ecosystem as evapotranspiration is smaller. On the other hand, boreal forests have less
919 evapotranspiration but they tend to consume a larger fraction of their precipitation
920 input. This finding is critical for understanding global patterns of forest productivity
921 and points to different adaptations to water supply where ecosystems accustomed to
922 an abundant supply of precipitation are not able to cope as well to dry anomalies.
923 Notably, boreal deciduous larch is a unique boreal functional type that differs from
924 evergreens due to distinct hydrological properties, including a heavier consumption of
925 precipitation and a better adaptation to the dryer and cooler climates. Comparison of
926 ecosystem-scale hydrologic properties across biomes shows vegetation adaptation for
927 the current climate conditions. Differences in ecosystem hydrologic properties may
928 impact their ability to adapt to short and long-term climate anomalies.

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