

# Mechanisms and effects of under-ice warming water in Ngoring Lake of Qinghai-Tibet Plateau

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**Abstract** The seasonal ice cover in lakes of the Qinghai-Tibet Plateau is a transient and vulnerable part of the cryosphere, whose characteristics depend on the regional climate: strong solar radiation in the context of the dry and cold environment because of high-altitude and relatively low-latitude. We use the first under-ice temperature observations from the largest Tibetan freshwater lake Ngoring and a one-dimensional lake model to quantify the mechanism of solar thermal accumulation under ice, which relies on the ice optical properties and weather conditions, as well as the effect of the accumulated heat on the land-atmosphere heat exchange after the ice break-up. The model was able to realistically simulate the feature of Ngoring Lake thermal regime: the “summer-like” temperature stratification with temperatures exceeding the maximum density point of 3.98 °C across the bulk of the water column. A series of sensitivity experiments revealed solar radiation was the major source of under-ice warming and demonstrated that the warming phenomenon was highly sensitive to the optical properties of ice. The heat accumulated under ice contributed to the heat release from the lake to the atmosphere for 1-2 months after ice-off, increasing the upward sensible and latent surface heat fluxes on average by  $\sim 50 \text{ W m}^{-2}$  and  $\sim 80 \text{ W m}^{-2}$ , respectively. Therefore, the delayed effect of heat release on the land-atmosphere interaction requires an adequate

34 representation in regional climate modeling of the Qinghai-Tibet Plateau and other  
35 lake-rich alpine areas.

36

## 37 **1 Introduction**

38 Seasonal lake ice is a part of the cryosphere, gaining recent attention from  
39 researchers due to its sensitivity to climate change (Kirillin et al., 2012; Sharma et al.,  
40 2020). The duration of ice cover affects the stability and vertical mixing of lakes, as  
41 well as the lake-atmosphere matter and energy exchange (Rösner et al., 2012; Efremova  
42 et al., 2013; Ramp et al., 2015). Ice cover regulates lake biochemical indicators, such  
43 as the concentration of dissolved oxygen, nitrogen, and phosphorus, changing the  
44 biochemical reaction rate and affecting the water quality and distribution of aquatic  
45 organisms (Weitere et al., 2010; Dokulil, 2013; Li et al., 2015a; Hardenbicker et al.,  
46 2016). Shortening of the ice season has been observed worldwide (Sharma et al., 2019;  
47 Dauginis and Brown, 2021) and attributed to anthropogenic warming (Grant et al.,  
48 2021). Future climate predictions indicated the accelerated reduction of seasonal lake  
49 ice, especially pronounced in the lake-rich Arctic regions (Brown and Duguay, 2011).  
50 Global assessment of seasonal lake ice changes requires quantification of the major heat  
51 sources and sinks on seasonal to climatic time scales. While the major prerequisite for  
52 the ice cover development is sufficient long season with air temperature below the  
53 freezing point of water, the heat budget of ice-covered lakes varies with latitude and  
54 altitude, depending strongly on the available solar radiation, the latter being the major  
55 source of heat for under-ice lake water (Kirillin et al., 2012). During the polar night in  
56 the Arctic and temperate lakes covered by snow, the solar heating is minor and the  
57 bottom sediment is the main heat source (Winter I according to Kirillin et al., 2012); at  
58 later stages of the ice season (Winter II), as the snow melts, solar radiation becomes to  
59 the main heat source governing thermal stratification and mixing under ice and the  
60 melting process at the ice base (Kirillin et al., 2018, 2020). Further, lakes with seasonal  
61 ice cover can be divided into cryomictic and cryostratified according to their maximum  
62 depth, surface area, and wind speed (Yang et al., 2021). In dry and cold areas with little  
63 snow, winter II can occupy the entire ice-covered period (Kirillin et al., 2012), making  
64 solar radiation to be the major factor affecting the lake ice regime. The situation is  
65 relevant to the alpine lakes.

66 In particular, the largest alpine lake system of the Qinghai-Tibet Plateau (TP), is  
67 not only the highest plateau on Earth with an average altitude of 4000-5000 m, but also  
68 located in the relatively low-latitude of 26-39° N ensures a high amount of solar  
69 radiation. The TP is covered by more than 1400 lakes with an area larger than 1 km<sup>2</sup>,

70 and the total lake area is more than  $5 \times 10^4$  km<sup>2</sup>, accounting for 57.2 % of that in China  
71 (Wan et al., 2016; Zhang et al., 2019). Recent studies reported the first observations  
72 from ice-covered Tibetan lakes, indicating the major role of solar radiation in their  
73 thermal regime (Wang et al., 2021). Water temperatures in Lakes Bangong Co and Nam  
74 Co constantly increased during the ice-covered period, with a stronger increase in  
75 shallower Bangong Co (Lazhu et al., 2021). Observations in meromictic Dagze Co Lake  
76 demonstrated stable temperature in the early ice-covered period start warming only in  
77 the late ice-covered period, conditioned by the high water salinity (Wang et al., 2014;  
78 Lazhu et al., 2021). Salinity has a strong influence on the temperature and mixing  
79 regime of all three abovementioned lakes, by altering their density stratification and  
80 vertical heat transport. Among freshwater lakes in the TP, Ngoring Lake is the largest  
81 one (Kirillin et al., 2017; Wen et al., 2022). Kirillin et al. (2021) found strong solar  
82 radiation under ice cover heating the entire lake water column to the maximum  
83 freshwater density temperature ( $\sim 3.98$  °C,  $T_{md}$ ) more than a month before the ice  
84 breakup—the situation never found in lowland freshwater lakes. As a result, strong heat  
85 release from water to the ice base turned into the major factor governing the ice melt,  
86 with the water temperature under ice achieving 6 °C. This radiation-dominated regime,  
87 differing dramatically from the typical heat budget known from earlier studies on ice-  
88 covered lakes. Although it belongs to the classification that winter II occupies the whole  
89 ice-covered period, the under-ice lake temperature exceeds 4 °C in the late stage due to  
90 the strong SR on the TP. Quantification of the resulting heat balance and thermal  
91 stratification characteristic of alpine conditions is the subject of the present study.

92 Due to the harsh environment of the TP and difficulties in collecting field  
93 observations, numerical models are often used to investigate phenomena and  
94 mechanisms of TP lakes. At present, the widely used lake models are the FLake model  
95 and the lake scheme coupled in the CLM (Community Land Model), CoLM (Common  
96 Land Model), and WRF (Weather Research and Forecasting Model) (Lazhu et al., 2016;  
97 Wen et al., 2016; Fang et al., 2017; Dai et al., 2018; Huang et al., 2019; Song et al.,  
98 2020; Wu et al., 2021). However, for computational efficiency, winter dynamics in these  
99 highly-parameterized lake models are represented in a rather simplified way, lacking  
100 the detailed mechanisms of heating by radiation and resulting vertical heat transports  
101 across the water column (Lazhu et al., 2016; Wen et al., 2016; Huang et al., 2019). As  
102 an alternative, we adopt for this study a “classical” two-equation turbulence modeling  
103 approach proving its reliability in decades of studies on the environmental turbulent  
104 fluid dynamics. The one-dimensional model LAKE implements the approach in  
105 application to lake dynamics and was applied previously to different lakes (Stepanenko  
106 et al., 2011, 2016; Guseva et al., 2016). We combine modeling with in situ observations  
107 from Ngoring Lake, data on weather forcing and remote sensing to: (i) test the ability

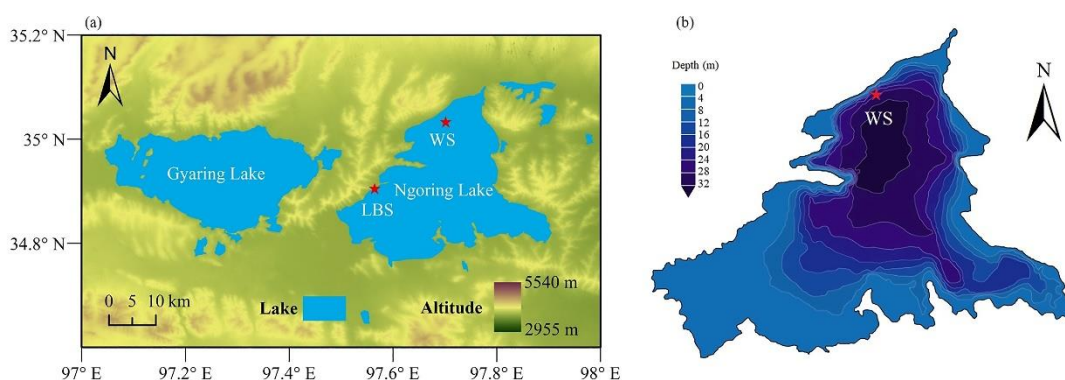
108 of a one-dimensional lake model LAKE to simulate temperature and stratification  
109 driven by intense solar heating in ice-covered Lake Ngoring; (ii) conduct series of  
110 sensitivity experiments aimed at revealing the role of meteorological forcing and ice  
111 optical properties in lake temperature and mixing regime; and (iii) reveal the effects of  
112 temperature distribution before ice breakup on lake heat storage and lake-atmosphere  
113 heat transfer.

114

## 115 2 Study area and data

### 116 2.1 Study area

117 Ngoring Lake (34.76-35.08° N, 97.53-97.90° E, Fig. 1) is located in the western  
118 valley of Maduo County on the eastern TP, with an average lake surface elevation of  
119 4274 m a.s.l. It is the largest freshwater lake in the Yellow River source region with a  
120 salinity of about 0.27 g kg<sup>-1</sup> (Shen et al., 2012). It has a surface area of 610 km<sup>2</sup>, a  
121 maximum depth of 32 m, and an average depth of 17 m. The pH is 8.49 and there are  
122 very few fish in the lake. Aquatic plants grow only in the riparian area. The lake  
123 thermally stratifies in summer and is covered by ice from late November or early  
124 December to late April (Wen et al., 2016). According to observational data from 1953  
125 to 2016 at Maduo station (34.9° N, 98.2° E) of the China Meteorological Administration,  
126 the average annual precipitation was 322.4 mm, mostly concentrated from May to  
127 September. The average annual air temperature was -3.53 °C. The maximum air  
128 temperature was 24.3 °C occurred on July 20, 2006, and the minimum air temperature  
129 was -48.1 °C occurred on January 2, 1978.



130

131 **Figure 1. (a) Location and (b) bathymetry of Ngoring Lake. (b) is adapted from**  
132 **Kirillin et al., 2021. The pentagrams denote the lake border station (LBS) and**  
133 **water temperature measurement site (WS).**

134

## 135 2.2 Data

### 136 2.2.1 Observations: LBS station and WS site

137 The long-term automatic lake border station (LBS, 34.91° N, 97.55° E, Fig. 1) was  
138 installed in October 2012, with an altitude of 4282 m a.s.l., providing meteorological  
139 forcing data: wind speed at 10 m, air temperature, specific humidity and air pressure at  
140 2 m, downward shortwave (SR) and longwave radiation (LR) at 1.5 m from September  
141 2015 to September 2016 (Li et al., 2020). The detailed information about site  
142 configuration and measured quantities are referred to in Li et al. (2015b) and Wen et al.  
143 (2016). The precipitation was obtained from the daily value data set (V3.0) at Maduo  
144 station of Chinese surface climate data (<http://data.cma.cn>).

145 The water temperature measurement site (WS, 35.03° N, 97.70° E, Fig. 1) was  
146 located in the northern of Ngoring Lake, where the total water depth was about 26.5 m.  
147 The multi-layer water temperature observation system consisted of 16 self-recording  
148 RBR SOLO water temperature probes with a precision of 0.01 °C. The sampling  
149 distance and time intervals were 1 m and 10 minutes, respectively.

150

### 151 2.2.2 MODIS lake surface temperature

152 The 8-day L3 global lake surface temperature product (MYD11C2) was derived  
153 from the data of Moderate Resolution Imaging Spectroradiometer (MODIS) and was  
154 used to evaluate the simulated results. MODIS offers long-term daily global coverage  
155 data with high spatial resolution. This product provides an 8-day combined radiative  
156 surface temperature at approximately 10:30 and 22:30 LT (local time), which is the  
157 satellite transit time. The resolution is 0.05° latitude/longitude (5600 m at the equator)  
158 for Climate Modeling Grid (CMG) (<https://ladsweb.nascom.nasa.gov/search>) (Wan et  
159 al., 2004).

160

### 161 2.2.3 ERA5-Land data

162 ERA5-Land is produced as an enhanced global dataset for the land component of  
163 the fifth generation of European ReAnalysis (ERA5) by the European Centre for  
164 Medium-Range Weather Forecasts (ECMWF), framed within the Copernicus Climate  
165 Change Service (C3S) of the European Commission. It is available for ERA5-Land  
166 hourly record for about 40 years from 1981 to the present. Expediently, ERA5-Land  
167 has an enhanced horizontal resolution of 9 km (~0.08°) compared to ERA5 (31 km) and  
168 ERA-Interim (80 km) (<https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis->

169 era5-land?tab=form) (Hersbach et al., 2020; Muñoz-Sabater et al., 2021).

170 ERA5-Land data is applied for a comparative analysis of warming mechanisms  
171 and thermal conditions in Tibetan ice-covered lakes against those in the Arctic. The  
172 reanalysis forcing data for the geographical position 69.05° N, 20.83° E was adopted as  
173 “typical” arctic weather conditions, that is, the SR is 0 W m<sup>-2</sup> during the winter polar  
174 night period. Northern Fennoscandia is covered by several lakes characterized by the  
175 longest ice-covered period in Western Europe. The largest of these lakes, Kilpisjärvi,  
176 has a similar morphometrical feature to Ngoring (average depth 19.5 m, maximum  
177 depth 57 m, surface area 37 km<sup>2</sup>). However, they receive different SR because of  
178 different latitudes. The lake has been intensively studied in the last decades (Kirillin et  
179 al., 2015, 2018; Leppäranta et al., 2017, 2019). Its under-ice water temperature  
180 remained stable during winter from 1992 to 1993 (Tolonen, 1998). In the following,  
181 model experiments forced by the ERA5 weather data (1992-1993) for the Arctic refer  
182 to “Kilpisjärvi” runs.

183

## 184 **3 Methods**

### 185 **3.1 LAKE model**

186 The one-dimensional model LAKE, simulating thermodynamic, hydrodynamic,  
187 and biogeochemical processes, is used to solve the horizontally averaged transfer of  
188 gases, heat, salts, and momentum in an enclosed water body (Stepanenko et al., 2011,  
189 2016). The vertical heat diffusion is simulated, and the penetration of solar radiation  
190 into the water ice, snow, and bottom sediments layers (Heiskanen et al., 2015; Cao et  
191 al., 2020) is taken into account. The exchange between the water and the inclined  
192 bottom is modeled explicitly because the model equations have been averaged over  
193 horizontal sections of the water body. The 2nd order  $\kappa$ - $\epsilon$  parametrization of turbulence  
194 is applied (Stepanenko et al., 2016).

195

#### 196 **3.1.1 Heat transfer in water body**

197 The water temperature is calculated according to the one-dimensional thermal  
198 diffusion equation:

$$199 \quad c_w \rho_w \frac{\partial T_w}{\partial t} = -c_w \rho_w \frac{1}{A} \int_{\Gamma_A} T_w (u_h \cdot n) dl + \frac{1}{Ah^2} \frac{\partial}{\partial \xi} \left( A_w K_T \frac{\partial T_w}{\partial \xi} \right) - \frac{1}{Ah} \frac{\partial AS}{\partial \xi} +$$
$$200 \quad \frac{1}{Ah} \frac{\partial A}{\partial \xi} [S_b(\xi) + F_{iz,b}(\xi)] + \frac{dh}{dt} \frac{\xi}{h} \frac{\partial T_w}{\partial \xi}, \quad (1)$$

201 where  $c_w$  is water specific heat,  $\rho_w$  is water density,  $T_w$  is water temperature,  $h(t)$  is

202 lake depth,  $t$  is time,  $\xi = z/h$  is a normalized vertical coordinate ( $z \in [0, h]$ ),  $z = 0$  is  
 203 located at the free water surface of the lake,  $S$  is downward shortwave radiation,  $A_w$  is  
 204 the  $z$ -dependent cross-sectional area of water,  $K_T$  is thermal diffusivity coefficient  
 205 equal to the sum of molecular and turbulent diffusivities,  $S_b(\xi)$  is shortwave radiation  
 206 flux,  $F_{i,z,b}$  is soil heat flux at the level  $z$ ,  $n$  is an outer normal vector to the boundary  $\Gamma_A$   
 207 of the horizontal cross-section  $A$  and  $u_h$  is horizontal vector in water (Stepanenko et al.,  
 208 2016; Guseva et al., 2016).

209

### 210 3.1.2 Heat transfer in ice cover

211 When the air temperature decreases below 0 °C and the surface water temperature  
 212 drops to the freezing point, the initial ice cover forms. When the net radiation of the  
 213 lake is positive, the ice melts continuously until the ice thickness declines to zero. The  
 214 general heat conduction equation in ice cover follows the equation:

$$215 \quad c_i \rho_i \frac{\partial T_i}{\partial t} = c_i \rho_i \frac{\xi}{h_i} \frac{dh_i}{dt} \frac{\partial T_i}{\partial \xi} - c_i \rho_i \frac{1}{h_i} \frac{dh_{i0}}{dt} \frac{\partial T_i}{\partial \xi} - \frac{1}{h_i} \frac{\partial S}{\partial \xi} + \frac{1}{A_i h_i^2} \frac{\partial}{\partial \xi} \left( A_i \lambda_i \frac{\partial T_i}{\partial \xi} \right) + \frac{1}{A_i h_i} \frac{\partial A_i}{\partial \xi} F_{T,b} -$$

$$216 \quad L \rho_i \frac{dp}{dt}, \quad (2)$$

217 where  $c_i$  is ice specific heat,  $\rho_i$  is ice density,  $T_i$  is ice temperature,  $\lambda_i$  is ice thermal  
 218 conductivity,  $h_i$  is ice thickness,  $\frac{dh_{i0}}{dt}$  is the increment of ice thickness on its surface,  
 219  $F_{T,b}$  is the heat flux at the ice-sediment boundary,  $A_i$  is the  $z$ -dependent cross-sectional  
 220 area of the ice cover determined by the basin morphometry,  $L$  is the latent heat of water  
 221 and  $p$  is ice porosity (Stepanenko et al., 2019). The last term to the right-hand side  
 222 presents heat of phase transition of salty water in ice pores.

223 The penetration of solar radiation into the medium is calculated using the Beer-  
 224 Lambert law (Stepanenko and Lykossov, 2005; Stepanenko et al., 2019):

$$225 \quad S(\xi) = S(0) \exp(-a_e h \xi), \quad (3)$$

226 where  $a_e$  is the medium extinction coefficient. To solve the temperature in Eq. (1 & 2),  
 227 it is necessary to specify the top and bottom boundary conditions and provide the  
 228 method to calculate the heat flux at each depth  $z$ . The atmospheric turbulent heat flux  
 229 schemes are based on the Monin-Obukhov similarity theory (Stepanenko et al., 2016).

230 When the lake is covered by ice, the temperatures of the bottom layer of ice and  
 231 the top layer of water are equal and fixed to the melting/freezing point temperature  
 232 (Stepanenko et al., 2019), which is calculated by the following formula:

$$233 \quad T_{mp} = -C * \left| \frac{\partial T_{mp}}{\partial C} \right|, \quad (4)$$

234 where  $T_{mp}$  is the melting/freezing point temperature (°C),  $C$  is salinity at the water-  
 235 ice interface,  $\left| \frac{\partial T_{mp}}{\partial C} \right| = 66.7$  is assumed constant.

236 Based on the study by Leppäranta (2014), the albedo regulates the surface energy  
 237 budget, and the extinction coefficient controls the vertical distribution of radiation  
 238 energy in the medium. In the LAKE model, the albedo of water ( $A_w$ ) is 0.06, and the  
 239 snow extinction coefficient ( $E_s$ ) decreases with increased snow density. Snow  
 240 accumulation in the Ngoring Lake area is almost zero. Therefore, only  $A_i$ ,  $E_i$ , and  $E_w$   
 241 are analyzed in this study. Version 2.3 called LAKE2.3 is used in this article.

242

### 243 **3.2 Methods to evaluate the model accuracy**

244 The indexes to evaluate accuracy of the model are the root mean square error  
 245 ( $RMSE$ ),  $BIAS$ , and correlation coefficient ( $CC$ ):

$$246 \quad RMSE = \sqrt{\frac{1}{n} \sum_{j=0}^n (m_j - o_j)^2} , \quad (5)$$

$$247 \quad BIAS = \bar{m} - \bar{o} , \quad (6)$$

$$248 \quad CC = \frac{\text{Cov}(M,O)}{\sqrt{\text{Var}(M)\text{Var}(O)}} , \quad (7)$$

249 where  $m_j$  and  $o_j$  represent the simulations and observations.  $\bar{m}$  and  $\bar{o}$  are the  
 250 corresponding average values.  $\text{Var}(M)$  and  $\text{Var}(O)$  are the variances of observed and  
 251 simulated values, respectively.  $\text{Cov}(M, O)$  is the covariances.

252

### 253 **3.3 Calculation method of heat storage**

254 The heat storage evolution in water is calculated by the following formulation:

$$255 \quad Q = c_w \rho_w \sum_{k=1}^n T_k \Delta z_k , \quad (8)$$

256 where  $c_w = 4192 \text{ J kg}^{-1} \text{ K}^{-1}$  and  $\rho_w = 10^3 \text{ kg m}^{-3}$ ,  $n$  is the layer number,  $\Delta z_k$  is depth  
 257 interval between two successive layers and  $T_k$  (K) is the average temperature in layer  $k$   
 258 (Nordbo et al., 2011; Gan and Liu, 2020).

259

## 260 **4 Characteristic analysis of water temperature and local climate in two lakes**

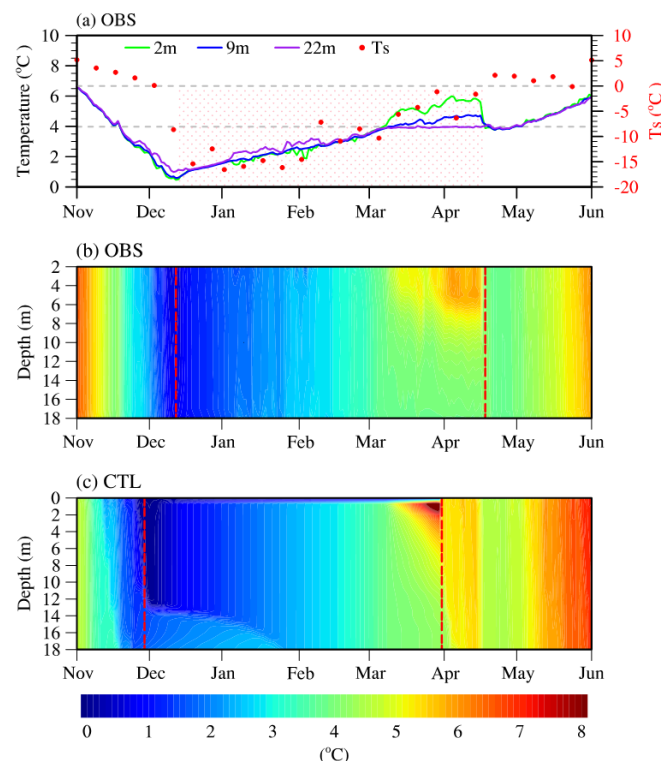
### 261 **4.1 Characteristics of water temperature**

262 It is pointed out that the under-ice water temperature from 2015 to 2016 in Ngoring  
 263 Lake rose continuously during the entire ice-covered period according to observations  
 264 (Wang et al., 2021; Kirillin et al., 2021). In November, because of the strong wind and



265 wide surface, the lake mixed evenly with slight oscillation ( $<1$  °C between 2 m and 22  
 266 m) and water temperature decreased gradually until the lowest point of 0.47 °C at 2 m  
 267 on December 12, the lake froze up completely (Fig. 2a). Meanwhile, the 2 m air  
 268 temperature fell to -7.79 °C. Ngoring Lake is mostly covered only by bare ice in winter  
 269 due to drought, less precipitation and snow. In the early ice-covered phase (from  
 270 December 12 to March 7), the whole lake mixed completely because solar radiation  
 271 penetrated ice and heated the upper water, which was warm ( $< T_{md}$ ), heavy and sinking  
 272 (Fig. 2b) (Kirillin et al., 2012). In parallel, water temperature continued to warm until  
 273 reached  $T_{md}$  on March 7 (Fig. 2a).

274 In the late ice-covered stage (from March 7 to April 18), the lake stratified. On the  
 275 one hand, owing to solar radiation strengthened, on the other hand, since radiation  
 276 absorption of water decayed with depth based on the Beer-Lambert law. Water  
 277 temperature increased at the rate of  $0.052$  °C  $d^{-1}$  in the layers from 2 m to 6 m, which  
 278 was more rapid than the early stage of  $0.035$  °C  $d^{-1}$ . On April 18, the ice melted entirely  
 279 as well water temperature rose to  $5.83$  °C at 2 m while remaining at  $T_{md}$  below 9 m.  
 280 After that, full mixing took place rapidly because the lake warmed gaining heat from  
 281 the sun and atmosphere as a result of ice breaking up (Fig. 2b).



282  
 283 **Figure 2. (a) The daily average water temperature observations of Ngoring Lake**  
 284 **at the surface ( $T_s$ ), 2 m, 9 m, and 22 m from November 1, 2015 to June 1, 2016.  $T_s$**   
 285 **is MODIS lake surface temperature. The gray reference lines denote 3.98 °C and**  
 286 **0 °C, respectively. The pink shaded area denotes ice-covered period. The water**

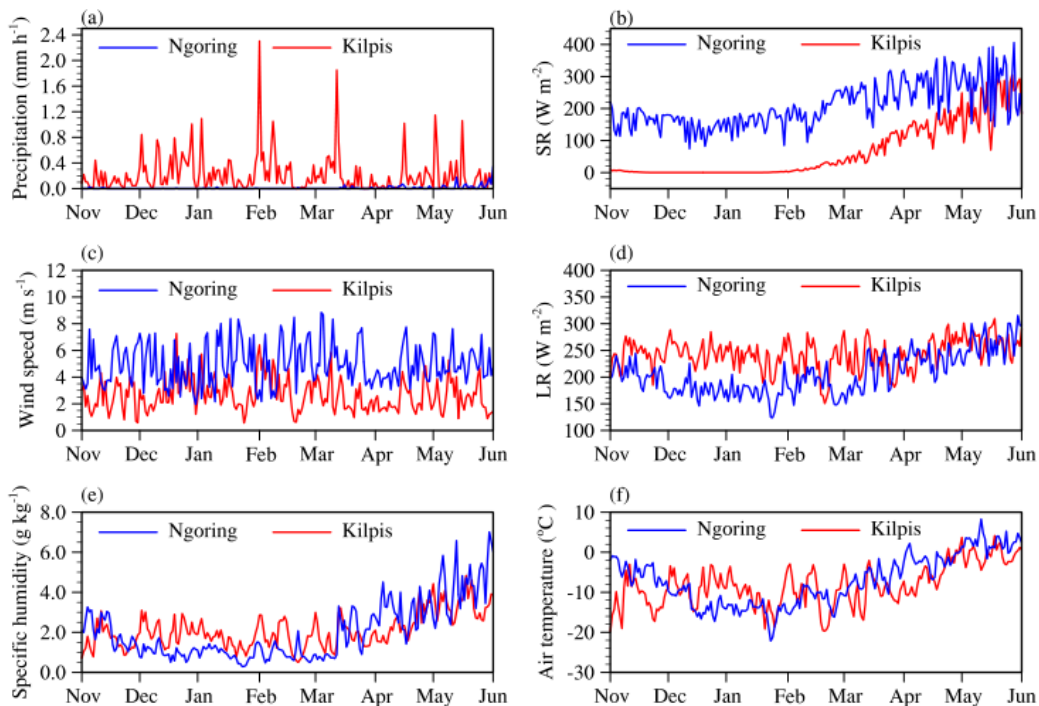
287 temperature profile (b) observed and (c) simulated in CTL. The ice-covered period  
 288 is represented between the two red dashed lines.

289

#### 290 4.2 Local climate: Tibet vs. Kilpisjärvi

291 The daily averages of meteorological variables between Tibet and Kilpisjärvi were  
 292 shown in Fig. 3, the ranges and averages of that during the ice-covered period (from  
 293 December 12 to April 18 based on Ngoring Lake) were compared (Table 1). The  
 294 average differences in air temperature, specific humidity, and downward LR were -  
 295 0.42 °C, -0.38 g kg<sup>-1</sup>, and 41.9 W m<sup>-2</sup>, respectively. The wind speed of Tibet was 1.7  
 296 times that of Kilpisjärvi, and the downward SR in Tibet of 199.41 W m<sup>-2</sup> was stronger  
 297 than in Kilpisjärvi of 40.46 W m<sup>-2</sup>. The precipitation was a multiple of 0.037 in Tibet  
 298 than that in Kilpisjärvi.

299 On the whole, there were few differences in air temperature, specific humidity, and  
 300 downward LR in the two regions. Nevertheless, there was much lower precipitation,  
 301 much higher downward SR and wind speed in the high-altitude and relatively low-  
 302 latitude Tibet. Surface pressure was not considered in this study since the little effect  
 303 on water temperature.



304

305 **Figure 3. Comparison of daily average values of the meteorological variables for**  
 306 **Tibet from 2015 to 2016 and for Kilpisjärvi from 1992 to 1993. (a) precipitation,**  
 307 **(b) downward SR, (c) wind speed, (d) downward LR, (e) specific humidity, and (f)**  
 308 **air temperature. The “Ngoring” represents the Tibet region and the “Kilpis”**  
 309 **represents Kilpisjärvi region.**

310 **Table 1.** Ranges and averages of the meteorological variables of Tibet (2015-2016)  
 311 compared with Kilpisjärvi (1992-1993) during the ice-covered period (12.12-4.18).

Meteorologic variables	Tibet		Kilpisjärvi	
	Range	Average	Range	Average
Precipitation (mm h <sup>-1</sup> )	< 0.072	0.0044	< 1.15	0.12
Downward SR (W m <sup>-2</sup> )	73.98-356.29	199.41	< 186.84	40.46
Wind speed (m s <sup>-1</sup> )	1.95-8.85	4.93	0.56-7.28	2.83
Downward LR (W m <sup>-2</sup> )	123.92-271.60	191.73	150.61-289.59	233.62
Specific humidity (g kg <sup>-1</sup> )	0.29-4.52	1.40	0.50-3.23	1.78
Air temperature (°C)	-22.16-2.24	-10.25	-19.69-2.01	-9.83

312

## 313 **5 Simulation setup**

314 To reveal the mechanism of water temperature rising during ice-covered period in  
 315 Ngoring Lake and its further influences, one control simulation (CTL) and 27  
 316 sensitivity simulations (SIM) depending on CTL were set in this study (Table 2).

317

### 318 **5.1 Setup in CTL**

319 The depth was set as 26.5 m, measured at the WS point, and divided vertically into  
 320 35 layers. The simulation period was from September 2015 to September 2016. The  
 321 initial vertical profile of water temperature, mixed layer, and the bottom temperature  
 322 were set following the observations (Fig. 2b). The albedos of snow and ice and the  
 323 extinction coefficients of ice and water were set as  $A_s = 0.7$ ,  $A_i = 0.25$ ,  $E_i = 2.5 \text{ m}^{-1}$ , and  
 324  $E_w = 0.15 \text{ m}^{-1}$  based on previous investigations (Lei et al., 2011; Li et al., 2018, 2020;

325 Shang et al., 2018). The input driving meteorological variables were air pressure, wind  
 326 speed, specific humidity, air temperature, precipitation, downward SR and LR. The  
 327 forcing data and model run intervals were 30 minutes and 15 seconds respectively.

328

## 329 5.2 Setup in SIM

330 To explore the influence of a single meteorological variable, SIM\_\* simulations  
 331 were set up. The symbol \* is SR, Precip, LR, U,  $T_{air}$ , or q in Kilpisjärvi. These scenarios  
 332 were quite artificial because these variables are closely correlated. Despite that, these  
 333 sensitivity simulations can shed light on the influence of local climate on lake  
 334 temperature evolution during ice-covered period.

335 To discuss the effect of main physical parameters, SIM\_# simulations were set up,  
 336 the sign # represented the values of  $A_i$ ,  $E_i$ , or  $E_w$ . SIM\_E\* (\* equal to 1, 2, or 3) is set  
 337 for exploring the effects of three different initial water temperature profiles before ice  
 338 breakup on the lake heat storage and heat fluxes.

339

340 **Table 2.** Names, explanations, and amounts of all experiments.

Experiment name	Experiment explanation	Amount
CTL	The control simulation	1
SIM_*		
(* represents meteorological variables)	The simulation when the * variable is replaced by that of Kilpisjärvi.	6
SIM_#		
(# represents values of $A_i$ , $E_i$ or $E_w$ )	The simulation when the corresponding physical variable is equal to #, respectively.	18
SIM_E*		
(* represents 1, 2, and 3)	The simulation when using three different initial temperature profiles before ice melting based on CTL.	3

341

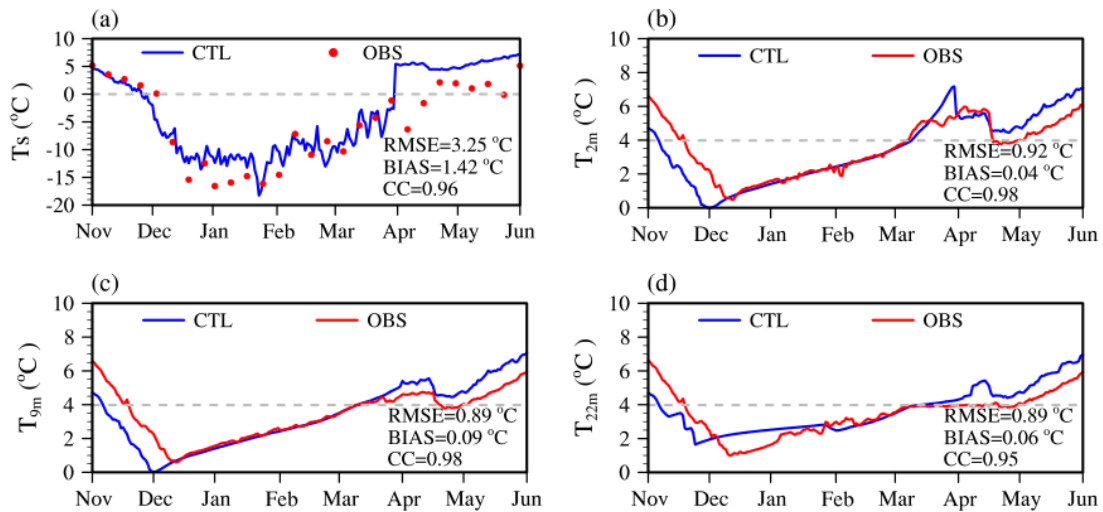
## 342 6 Simulation results

### 343 6.1 Model validation

344 The CTL simulation was relatively consistent with the observations (Fig. 2b,c),

345 even though the whole ice season advanced by about 15 days than observed. Water  
 346 temperature was a little higher in CTL than that in the observations from mid-March to  
 347 late May. The deviation was greater in the deep water. The simulated temperature  
 348 warmed faster and higher by 1 °C than the observed value after ice melted.

349 The results were evaluated by calculating *RMSE*, *BIAS*, *CC* between simulated and  
 350 observed water temperature at the lake surface (*T<sub>s</sub>*), 2 m, 9 m and 22 m (Fig. 4). The  
 351 *CC* in each layer was equal to even greater than 0.95, and the *CC* in 2 m and 9 m are as  
 352 high as 0.98, even though *RMSE* and *BIAS* of lake surface were 3.25 °C and 1.42 °C,  
 353 respectively. The *T<sub>s</sub>* *RMSE* was largely due to the uncertainty of MODIS lake surface  
 354 temperature (Donlon et al., 2002; Tavares et al., 2019). The *BIAS* absolute values in the  
 355 internal lake were less than 0.01 °C, and *RMSE* was less than 0.95 °C. More important,  
 356 the under-ice water temperature warming phenomenon was reproduced reasonably.



357

358 **Figure 4. The daily average water temperature observed and simulated in CTL of**  
 359 **(a) the surface (*T<sub>s</sub>*), (b) 2 m, (c) 9 m, (d) 22 m in Ngoring Lake from November**  
 360 **2015 to June 2016. The dotted lines represent *T<sub>md</sub>* 3.98 °C.**

361

## 362 6.2 Influences of local climate on water temperature

363 To explore the influences of local climate on water temperature, six simulations  
 364 SIM\_\* (\* represents 6 meteorological variables, Table 2) were designed. The 3 m water  
 365 temperature was typically selected to analyze since water temperature at different  
 366 depths varied consistently over time.

367 SIM\_SR was the simulation when the Kilpisjärvi downward SR was substituted  
 368 for that in Tibet. During the ice-covered period, the downward SR difference between  
 369 CTL (199.41 W m<sup>-2</sup>) and SIM\_SR (40.46 W m<sup>-2</sup>) was 158.95 W m<sup>-2</sup>. In the sensitivity  
 370 simulation SIM\_SR, the 3 m water temperature was stable keeping in the range of 0-

371 0.1 °C (Fig. 5a). The ice formation date was earlier and the ice-breaking date delayed,  
372 which led to the growth of the whole ice season. The mixed layer depth increased (Fig.  
373 5d). Consequently, the strong downward SR on the TP generated the under-ice water  
374 warming in Ngoring Lake.

375 In the simulation SIM\_Precip, the Tibet precipitation was replaced by that in  
376 Kilpisjärvi. In the sensitivity experiment SIM\_Precip, the 3 m water temperature fixed,  
377 then increased but did not exceed  $T_{md}$  in the early ice cover stage (Fig. 5a). The  
378 stratification and temperature maximum center disappeared in late March, and the lake  
379 was fully mixed (Fig. 2c,5g). Because the average precipitation in SIM\_Precip (0.12  
380 mm h<sup>-1</sup>) was approximately 30 times larger than that in CTL (0.0044 mm h<sup>-1</sup>) during  
381 ice-covered period, more solar radiation was reflected and absorbed by snow due to  
382 more snowfall accumulation. Thus, the high precipitation damped the water  
383 temperature rise.

384 In SIM\_LR simulation, the downward LR in Kilpisjärvi superseded that in Tibet.  
385 The average downward LR was 233.62 W m<sup>-2</sup> in SIM\_LR, which was stronger than  
386 that in CTL (191.73 W m<sup>-2</sup>) during ice-covered period. The 3 m water temperature still  
387 warmed as well the complete ice melting time of late February was ahead. The heat was  
388 transferred from lake to atmosphere because of lower air temperature after ice breakup.  
389 The water temperature underwent a cooling process (2 °C) until reaching a new  
390 equilibrium with atmosphere (Fig. 5b). Compared with the CTL, water mixing in the  
391 ice-covered period was more uniform, the stratification in late March was weakened,  
392 and the temperature maximum center advanced about 15 days (Fig. 5e).

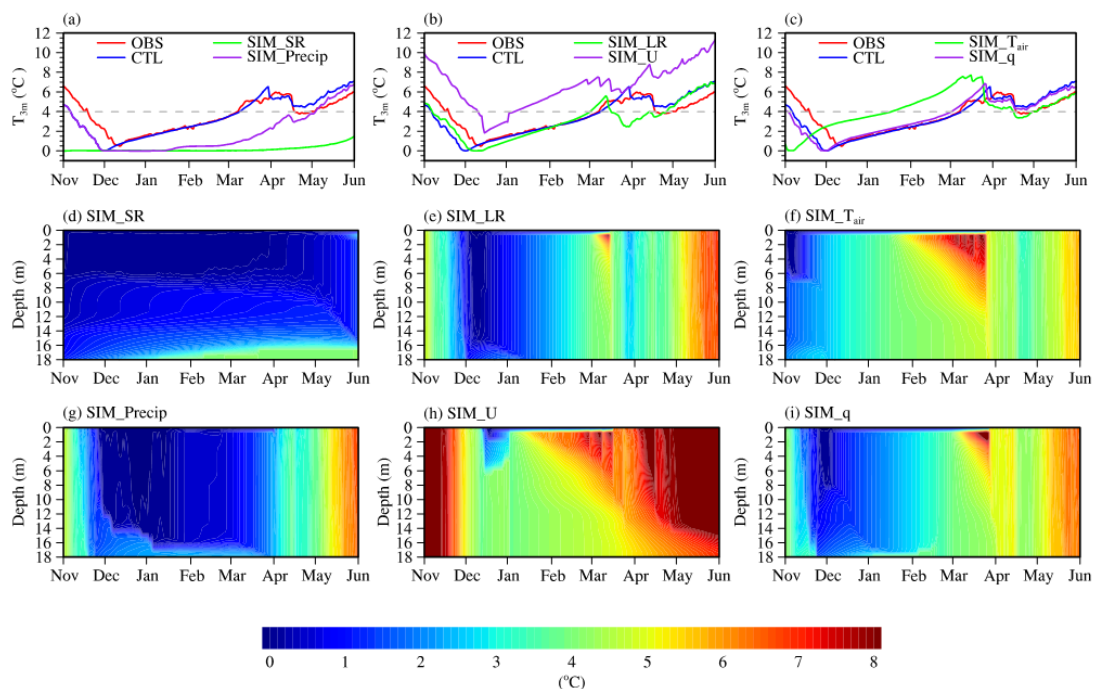
393 In SIM\_U simulation, the wind speed of Kilpisjärvi was substituted for that of Tibet.  
394 The wind speed in SIM\_U (2.83 m s<sup>-1</sup>) was weaker than that in CTL (4.93 m s<sup>-1</sup>) for the  
395 ice-covered period. In the sensitivity experiment SIM\_U, the 3 m water temperature  
396 kept rising, but it was about 3 °C higher than that in CTL during the whole simulation  
397 period (Fig. 5b). Due to the decrease in wind speed, the mixed layer depth was reduced,  
398 and the lake stratification was more stable (Fig. 5h).

399 In SIM\_T<sub>air</sub> simulation, the air temperature of Kilpisjärvi was replaced by that of  
400 Tibet. The average air temperature difference between SIM\_T<sub>air</sub> (-9.83 °C) and CTL (-  
401 10.25 °C) was negligible (0.42 °C). In the sensitivity experiment SIM\_T<sub>air</sub>, the water  
402 temperature decreased more quickly, and in late October, the lake froze no longer  
403 releasing heat into atmosphere. The lake stratification was enhanced, and the water  
404 temperature maximum center was ahead about 10 days (Fig. 5c,f).

405 In SIM\_q simulation, the specific humidity of Kilpisjärvi was substituted for Tibet.  
406 The difference in specific humidity between SIM\_q and CTL was 0.38 g kg<sup>-1</sup> during

407 ice-covered period. In the sensitivity experiment SIM<sub>q</sub>, the simulations were  
 408 coincidental to that in CTL, and thus the specific humidity had little effect on the water  
 409 temperature (Figs. 5c,i).

410 In conclusion, the stronger downward SR and lower precipitation in the **high-**  
 411 **altitude and relatively low-latitude** TP played positive roles in the water temperature  
 412 warming during the ice-covered period in Ngoring Lake. Less downward LR, lower air  
 413 temperature, and larger wind speed did not change the warming trend but affected the  
 414 warming amplitude and rate. Specific humidity had no significant influence.



415

416 **Figure 5. The simulated 3 m daily average water temperature in (a) (d) SIM\_SR,**  
 417 **(a) (g) SIM\_Precip, (b) (e) SIM\_LR, (b) (h) SIM\_U, (c) (f) SIM\_T<sub>air</sub>, (c) (i) SIM\_q**  
 418 **sensitivity experiments from November 2015 to June 2016 are compared with the**  
 419 **CTL and the observation, and the change of vertical stratification is shown. The**  
 420 **dotted line represents 3.98 °C.**

421

### 422 6.3 Influences of main physical parameters on water temperature

423 The radiation transfer, which depended on the albedo and extinction coefficient,  
 424 played a decisive role in the water temperature. Only influences of  $A_i$ ,  $E_i$ , and  $E_w$   
 425 on water temperature simulation were discussed with sensitivity experiments due to less  
 426 snow in Tibet. According to previous observations,  $A_i$  has observed on TP was mostly  
 427 less than 0.12, and the albedo of clear blue ice was only 0.075 (Li et al., 2018). The  
 428 range of  $A_i$  without snow cover was set as 0.1-0.8 with an interval of 0.1 in SIM\_ $A_i$

429 experiments.

430  $E_i$  has not been observed on TP, but surveys in Finnish lakes show that the value of  
431 bare ice varies between 1-4  $\text{m}^{-1}$ , while the value of snow-covered ice can reach 5  $\text{m}^{-1}$   
432 (Lei et al., 2011). In SIM\_ $E_i$  simulations,  $E_i$  was equal to 1-5  $\text{m}^{-1}$  with an interval of 1  
433  $\text{m}^{-1}$ .

434 For the  $E_w$ , Zolfaghari et al. (2017) found that the FLake model is particularly  
435 sensitive at  $E_w \leq 0.5 \text{ m}^{-1}$ . Shang et al. (2018) observed that  $E_i$  varies from 0.11 to 0.67  
436  $\text{m}^{-1}$  in a few TP lakes. Therefore, the sensitivity simulations SIM\_ $E_w$  were designed in  
437 which the  $E_w$  varied from 0.1 to 0.5  $\text{m}^{-1}$  with an increment step of 0.1  $\text{m}^{-1}$ . The  
438 experimental settings are shown in Table 3.

439

440 **Table 3.** Numerical sensitivity simulations of parameters affecting the radiative transfer.

Parameter	CTL	SIM_ $A_i$	SIM_ $E_i$	SIM_ $E_w$
$A_i$	0.25	0.1/0.2/0.3/0.4/0.5 /0.6/0.7/0.8	0.25	0.25
$E_i$ ( $\text{m}^{-1}$ )	2.5	2.5	1.0/2.0/3.0/4.0 /5.0	2.5
$E_w$ ( $\text{m}^{-1}$ )	0.15	0.15	0.15	0.1/0.2/0.3/0.4 /0.5

441

442 In the SIM\_ $A_i$  sensitivity experiment, the 3 m water temperature decreased  
443 approximately 1 °C when ice albedo increased by 0.1. When the albedo grew to 0.80,  
444 the water temperature warming decreased from 4 °C to 2 °C. The increase of ice albedo  
445 did not affect the ice formation date but remarkably delayed the ice melting time,  
446 accordingly prolonging the ice-covered period. When the albedo increased from 0.1 to  
447 0.8, the ice-covered period was extended for 15-30 days for every 0.1 increase (Fig. 6a).

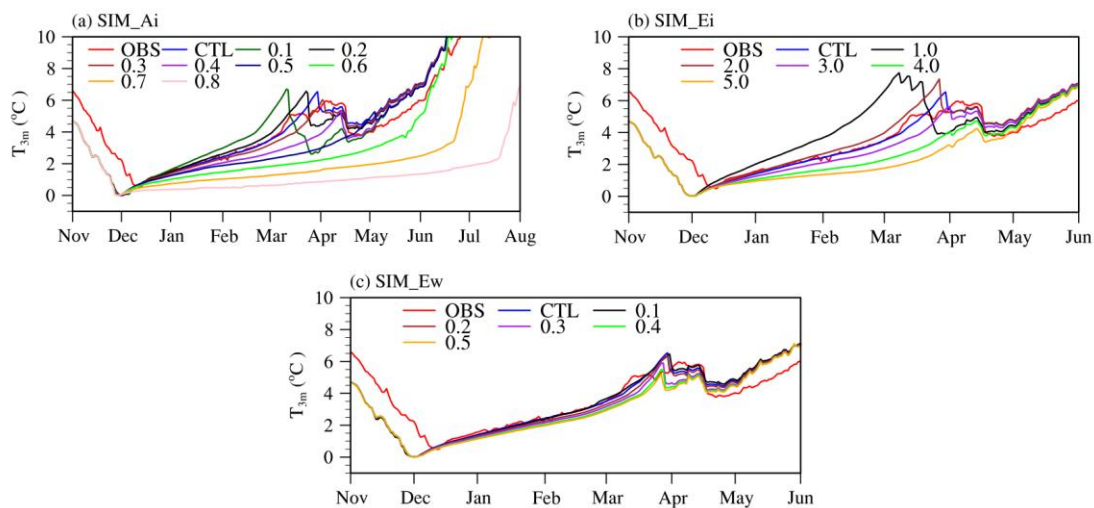
448 In the sensitivity experiment SIM\_ $E_i$ , the ice extinction coefficient changes did not  
449 all make a continuous rising in water temperature, but the 3 m water temperature  
450 decreased by 1-2 °C when ice extinction coefficient increased by 1  $\text{m}^{-1}$  (Fig. 6b). The  
451 ice absorbed more heat, and the less heat entered the lake water under ice due to the  
452 larger ice extinction coefficient.

453 To further discuss influences of  $A_i$  and  $E_i$  on lake temperature during ice-covered  
454 period. The period was divided into two stages Period-A and Period-B in CTL, SIM\_ $A_i$ .



455 Period-A ranged from freezing point to  $T_{md}$ , Period-B ranged from  $T_{md}$  to maximum  
 456 temperature ( $T_m$ ). The duration of Period-A is longer than that of Period-B, and the  
 457 temperature heating rate in Period-B ( $\sim 0.1 \text{ }^\circ\text{C d}^{-1}$ ) was 2.5 times greater than that of  
 458 Period-A ( $\sim 0.04 \text{ }^\circ\text{C d}^{-1}$ ). The reason was that lake completely covered by ice, and the  
 459 inner lake evenly mixed in Period-A, while the ice thickness decreased and the radiation  
 460 absorbed by the ice decreased in Period-B. The upper layer absorbed more heat than  
 461 the deeper layer, and the upper water temperature increased rapidly. When  $A_i$  and  $E_i$   
 462 increased, the heating rate decreased and the duration increased in Period-A, the  
 463 maximum temperature decreased, the heating rate and duration fluctuate in Period-B.  
 464 When  $A_i \geq 0.6$ , the heating rate during ice-covered period decreases and did not rise to  
 465  $T_{md}$ .

466 In the SIM\_Ew sensitivity experiment, the water extinction coefficient had just a  
 467 little influence on the winter water temperature, 3 m water temperature decreased with  
 468 the increase of  $E_w$  (Fig. 6c). When only the extinction coefficient of water changed, the  
 469 solar radiation entering the water through the ice is unchanged, and so the heat storage  
 470 of the lake was unaffected. It is just that the heat distribution in the vertical direction is  
 471 changed. The higher the extinction coefficient of water, the more heat was absorbed by  
 472 the surface layer and the less heat reached the deep layer. The phenomenon that the 3  
 473 m water temperature decreases with the extinction coefficient increasing becomes more  
 474 and more obvious in the later stage of ice melting.



475

476 **Figure 6. Comparison of the 3 m simulated daily average water temperature with**  
 477 **the observed value under different (a)  $A_i$ , (b)  $E_i$ , (c)  $E_w$ .**

478

#### 479 **6.4 Influences of water temperature on lake-atmosphere exchange**

480 The thermal conditions in an ice-covered lake just before ice melting have a

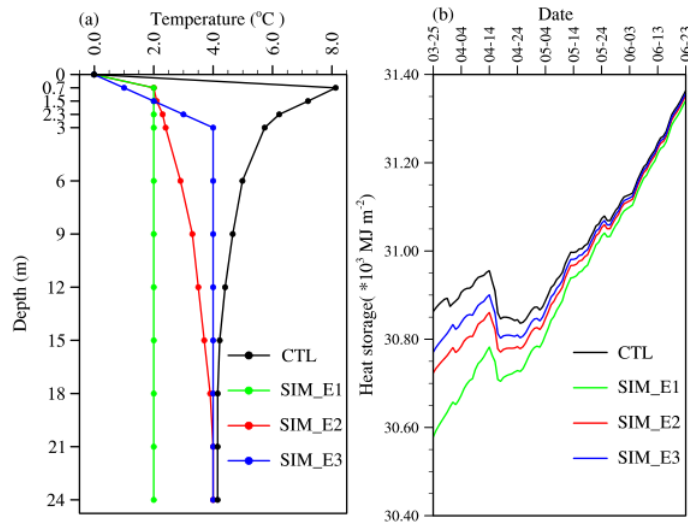
481 significant influence on the air-lake energy exchange. To analyze the effects of lake  
482 temperature characteristics on the atmosphere at ice melting, three experiments –  
483 SIM\_E1, SIM\_E2, and SIM\_E3 (Table 1) – were set up based on the CTL and the  
484 observed lake temperature profile on March 25, 2016, 5 days before ice completely  
485 melted (Fig. 7a). The characteristics of the initial water temperature profile were:

- 486 - SIM\_E1. The stratification was weak, the first layer temperature was at the  
487 melting point, and, from the second layer down, the water temperature was set  
488 as 2 °C corresponding to Bangong Co (Wang et al., 2014).
- 489 - SIM\_E2. The lake was strongly stratified. The first layer was at the melting  
490 point, and the temperature increased linearly reaching  $T_{md}$  at the bottom,  
491 corresponding to Valkea-Kotinen Lake (Bai et al., 2016).
- 492 - SIM\_E3. The temperature of the first layer was at the melting point, and the  
493 temperature gradually increased with the depth from the second layer to the  
494 middle layer, and the temperature in the middle layer increased to  $T_{md}$   
495 corresponding to Thrush Lake (Fang and Stefan, 1996).

496 In CTL, the first layer temperature equal to the freezing/melting point, and the  
497 second layer reached the maximum temperature on March 25. The temperature became  
498 lower with the deeper layer, until the temperature reached  $T_{md}$ .

499 With the different initial temperature profiles, the heat storage was different after  
500 ice breakup, and the difference persisted for about two months (Fig. 7b). In CTL, from  
501 March 30 to March 31 when ice melted completely, the lake heat storage ranged from  
502 30893.02 MJ m<sup>-2</sup> to 30874.51 MJ m<sup>-2</sup>, and the heat released was 18.51 MJ m<sup>-2</sup>. In the  
503 three experiments, from April 1 to April 2 when ice melted completely, the lake heat  
504 storage changed from 30657.51 MJ m<sup>-2</sup> to 30651.67 MJ m<sup>-2</sup> in SIM\_E1, from 30781.07  
505 MJ m<sup>-2</sup> to 30769.91 MJ m<sup>-2</sup> in SIM\_E2, and from 30833.28 MJ m<sup>-2</sup> to 30822.42 MJ m<sup>-2</sup>  
506 in SIM\_E3, and the heat release was 5.84 MJ m<sup>-2</sup>, 11.16 MJ m<sup>-2</sup>, and 10.86 MJ m<sup>-2</sup>,  
507 respectively (Fig. 7b).

508 The heat released was in the form of sensible heat and latent heat, accounting for  
509 0.060% (CTL), 0.019% (SIM\_E1), 0.036% (SIM\_E2), and 0.035% (SIM\_E3) of the  
510 ice-covered heat storage, respectively. As the initial lake temperature profiles were  
511 different before ice complete melting, the ice melted earlier and faster with the higher  
512 lake temperature. The lake heat storage increased from March 25 to May 24, and the  
513 heat release rate was different under different circumstances. After late May, the heat  
514 balance between the lake and the atmosphere was the same, and so the heat storage  
515 basically stayed equally after that.

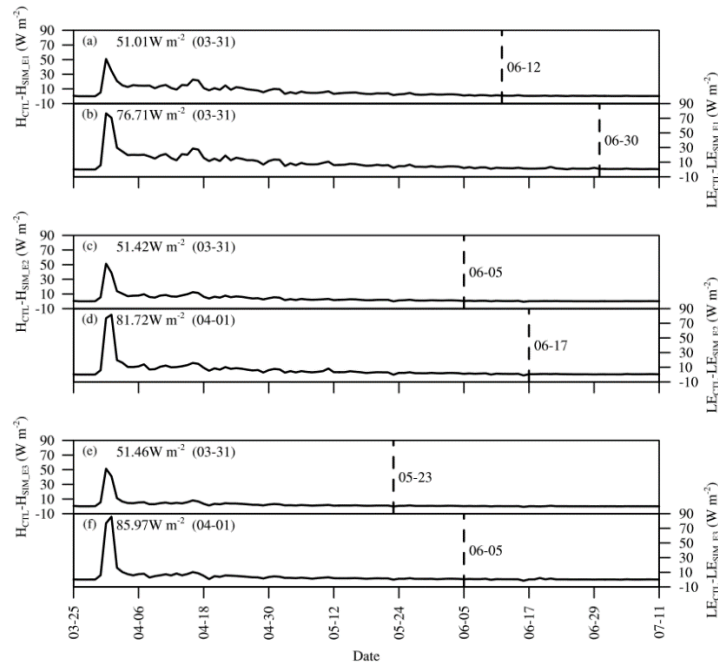


516

517 **Figure 7. (a) The initial water temperature profiles in the model are set on March**  
 518 **25, 2016, and the corresponding daily average (b) lake heat storage is simulated.**  
 519 **SIM\_E1, SIM\_E2, and SIM\_E3 are three different sensitivity simulations.**

520

521 The lake surface temperature also affected the sensible and latent heat release,  
 522 whose differences were calculated between CTL and the three experimental simulations  
 523 (Fig. 8). The influence of different initial water temperature profiles started on March  
 524 31, that is, when the ice had melted completely in CTL, and when the sensible and latent  
 525 differences between CTL and three experimental simulations were less than  $0.1 \text{ W m}^{-2}$   
 526 for three consecutive days, we judged that the influence had ended. The maximum  
 527 differences of the sensible heat ( $51.0 \text{ W m}^{-2}$ ) and latent heat ( $76.7 \text{ W m}^{-2}$ ) between  
 528 SIM\_E1 and CTL appeared on March 31 and ended on June 12 and 30, respectively  
 529 (Fig. 8a). In SIM\_E2 the corresponding numbers were  $51.4 \text{ W m}^{-2}$  (March 31 to June  
 530 5) for sensible heat and  $81.7 \text{ W m}^{-2}$  (April 1 to June 17) for latent heat (Fig. 8b), and in  
 531 SIM\_E3 they were  $51.5 \text{ W m}^{-2}$  (March 31 to May 23) for sensible heat and  $86.0 \text{ W m}^{-2}$   
 532 (April 1 to June 5) for latent heat (Fig. 8c). Compared with the three lake temperature  
 533 experiments, the heating characteristics of Ngoring Lake made the heat release higher  
 534 and faster during ice breakup. The duration of heat release difference was from 59 (to  
 535 May 23) to 97 (to June 30) days, and for the latent heat release, the situation lasted  
 536 about 12-18 days longer than for the sensible heat release.



537

538 **Figure 8. The daily average difference between the sensitivity sensible and latent**  
 539 **heat and the CTL under three different initial water temperature profiles in**  
 540 **SIM\_E1, SIM\_E2, and SIM\_E3.**

541

## 542 7 Conclusions

543 The analysis demonstrates a significant increase in lake temperature during the ice-  
 544 covered period in Ngoring, the largest freshwater lake on the Tibetan Plateau (TP), with  
 545 water temperatures exceeding the freshwater maximum density value  $T_{md}$ . The heating  
 546 is governed by strong solar radiation, the factor differing alpine lakes on the **high-**  
 547 **altitude and relatively low-latitude** TP from the **high-latitude and** low-altitude northern  
 548 lakes with similar winter air temperature patterns. The one-dimensional lake model  
 549 LAKE2.3 successfully captured the major mechanisms of warming and vertical thermal  
 550 stratification during the ice-covered period. Compared with MODIS surface  
 551 temperature data, the *BIAS*, *RMSE*, and *CC* were 1.42 °C, 3.25 °C, and 0.96,  
 552 respectively. The absolute values of *BIAS* and *RMSE* were less than 0.1 °C and 1 °C in  
 553 2 m, 9 m, 14 m, and 22 m. The *CC* of simulated and observed water temperature at 2  
 554 m, 9 m, and 14 m were as high as 0.98, and the *CC* of simulated and observed water  
 555 temperature at 22 m was 0.95.

556 Sensitivity simulations with perturbed local climate data confirmed the decisive  
 557 role of subsurface solar radiation in the water temperature rise and demonstrated strong  
 558 negative feedback with winter precipitation amount. The downward longwave radiation,  
 559 air temperature, and wind speed had only a minor influence on the water temperature.

560 The warming rates and, as a result, the duration of the ice-covered period was  
561 sensitive to the physical ice properties: ice albedo and light extinction coefficient both  
562 reduced the amount of the subsurface solar radiation. An increase in the albedo of ice  
563 reduced the rising trend of water temperature and prolonged the ice season. At the  
564 critical albedo of 0.6, the lake water warming decreased obviously and temperature  
565 remained stable no more than 3.98 °C. The extinction coefficient of water had just a  
566 minor effect on water temperature under the ice.

567 An important consequence of the under-ice solar heat accumulation consisted in  
568 increased sensible and latent heat releases in the subsequent open-water phase.  
569 According to the model results, the effects on the surface fluxes of Ngoring Lake lasted  
570 for 59-97 days after the ice melt and increased the upward latent and sensible surface  
571 heat fluxes up to  $\sim 80 \text{ W m}^{-2}$  and  $\sim 50 \text{ W m}^{-2}$ , respectively. Herewith, the phenomenon  
572 of under-ice solar heating may have a significant effect on the land-atmosphere  
573 interaction on regional scales and has to be accounted for in coupled climate models.

574

575 *Data availability.* The daily precipitation data from Chinese surface stations are  
576 available for purchase from the China Meteorological Data Service Center (CMDC,  
577 <http://data.cma.cn/en/>). The MODIS LST product is available from National  
578 Aeronautics and Space Administration (NASA) (<https://earthdata.nasa.gov/>). ERA5-  
579 Land data is available with funding from the European Union's Copernicus Climate  
580 Change Service (<https://cds.climate.copernicus.eu/>). Lake temperature data of Ngoring  
581 Lake in 2015 and 2016 were uploaded to Zenodo by Georgiy Kirillin  
582 (<http://doi.org/10.5281/zenodo.4750910>). The weather observation data of Ngoring  
583 Lake can be obtained from the website (<https://nimbus.igb-berlin.de/index.php/s/Moqxgn29DbNFyr8>). The latest version of LAKE model source  
584 code is available at zenodo: <https://zenodo.org/record/6353238#.YjCSXi1eNTY>.

586

587 *Author contributions.* MW and LW conceived the study. MW performed the modeling  
588 with contributions from VS, LW, and ZL. YZ, RN, and LY processed some data. MW,  
589 LW, ML, and GK analyzed the model output. MW wrote the paper, with contributions  
590 from all co-authors.

591

592 *Competing interests.* The authors declare that they have no conflict of interest.

593

594 *Acknowledgments.* This study was supported by the National Key Research and  
595 Development Program of China (2019YFE0197600) and CAS "Light of West China"  
596 Program (E129030101, Y929641001). Victor Stepanenko was supported by Russian  
597 Ministry of Science and Higher Education, agreement No. 075-15-2019-1621.

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