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EVAPORATION AND TRANSPORT OF WATER ISOTOPOLOGUES
FROM GREENLAND LAKES: THE LAKE SIZE EFFECT

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ABSTRACT

Isotopic compositions of evaporative flux from a lake are used in many hydrological and paleoclimate studies that help constrain the water budget of a lake and/or to infer changes in climate conditions. The isotopic fluxes of evaporation from a water surface are typically computed using a zero dimensional (0-D) model originally conceptualized by Craig and Gordon (1965). Such models generally have laminar and turbulent layers, assume a steady state condition, and neglect horizontal variations. In particular, the effect of advection on isotopic variations is not considered. While this classical treatment can be used for some sections of large open surface water bodies, such as an ocean or a large lake, it may not apply to relatively small water bodies where limited fetch does not allow full equilibration between air from land and the water surface. Both horizontal and vertical gradients in water vapor concentration and isotopic ratios may develop over a lake. These gradients, in turn, affect the evaporative fluxes of water vapor and its isotopic ratios, which is not adequately predicted by a 0-D model.

We observed, for the first time, the vertical as well as horizontal components of vapor and isotopic gradients as relatively dry and isotopically depleted air advected over the surfaces of several lakes up to 5 km fetch under winds of 1-5 m/s in Kangerlussuaq, Greenland. We modeled the vapor and isotopic distribution in air above the lake using a steady state 2-D model, in which vertical diffusive transport balances horizontal advection. The model was verified by our observations, and then used to calculate evaporative fluxes of vapor and its isotopic ratios. In the special case of zero wind speed, the model reduces to 1-D. Results from this 1-D model are compared with those from the 2-D model to assess the discrepancy in isotopic fluxes between advection and no advection conditions.

Since wind advection above a lake alters the concentrations, gradients, and evaporative fluxes of water isotopes, it alters the water balance and isotope ratios of the lake and the relationship between them. These effects are greatest for small lakes. If wind advection is neglected in the inference of water balance from lake isotopes, an error is thus introduced, the magnitude of which depends on lake size. We refer to this as the “lake size effect”. For lakes less than 500 m in length along the wind direction, the average $\delta^{18}\text{O}$ and δD of vapor flux are at least 2‰ lower than the corresponding flux values from the 1-D model. The magnitude of the resulting error in water balance calculations is much greater if using $\delta^{18}\text{O}$ than δD in mass balance calculations; the former is about eight times the latter. This result argues that water balance calculated with δD is less sensitive to the difference in lake size and/or its change over time.

76 The 1-D model result is also compared with that from a comparable 0-D model.
77 Since vertical vapor and isotope gradients always exist (even under no advection
78 conditions), one may not obtain correct flux values if the relative humidity and isotopic
79 ratios in ambient air measured at an arbitrary height are used for the 0-D model
80 calculation. Typically, the standard meteorological measurements at 2 or 10 m would
81 result in an underestimate of the $\delta^{18}\text{O}$ and δD values of the vapor flux.

82 This work has provided the first quantification on the effect of advection on
83 isotopic fluxes of evaporation. The method of mobile vapor analysis combined with 2-D
84 modeling can be applied to other environmental settings, in which the size of advection
85 effect on isotopic fluxes depends upon relationships among local meteorological and
86 hydrological variables. Our results also suggest that incorporating isotopic vapor
87 measurements can help constrain modeled evaporation rates, which is worth exploring
88 further in future studies.

89 **1. INTRODUCTION**

90 Evaporation is a vital component of the hydrological cycle, and constraints on
91 fluxes of vapor, heat and isotopologues associated with evaporation are important for
92 the quantification of many processes related to the global water cycle, regional water
93 balance, and interpretation of paleo-isotopic records. In Arctic regions, where lakes may
94 cover as much as 30% of terrestrial surface areas (Walter et al., 2006), lake evaporation
95 represents a significant fraction of surface water flux. Knowledge of this flux under
96 Arctic conditions is thus important to understanding the hydrological response under
97 rapidly changing climate in the polar environment (Zhang et al. 2011; Bourassa et al.
98 2012). Changes in water balance are also inferred from paleo-lake records, which reflect
99 changes of climate variables, such as precipitation and evaporation. Quantifying
100 evaporative fluxes is thus a critical part of paleoclimate studies that are based on lake
101 sediment records. This study focuses on evaporation of Arctic lakes.

102 Stable isotopic methods have been widely used to study lake evaporation or to
103 quantify water balance in which lake evaporation represents an important flux (e.g.,
104 Gonflantini 1986; Gibson et al. 2002; Gat et al. 1994). Understanding lake evaporation is
105 also important for interpreting isotopic records from lake sediments that record isotopic
106 compositions of lake water as an important environmental indicator (Sauer et al. 2001,
107 Sauer et al. 2001, Zhang and Sachs 2007; Jones and Imbers, 2010; Steinman and Abbott,
108 2013). One important variable that controls evaporative isotopic fluxes is the water
109 vapor isotopic composition above an evaporative surface (Craig and Gordon 1965). This
110 measurement has not been easy until recent developments in spectrum based laser

111 technology (e.g., Gupta et al. 2009). Our project takes advantage of this new
112 technology's ability for rapid data acquisition and mobile deployment.

113 In most previous studies, the evaporative isotopic enrichment of lake water is
114 typically quantified by a zero dimensional (0-D) steady state model (e.g., Gibson et al.,
115 1993; Yi et al., 2007; Turner et al., 2014). Craig and Gordon (1965) first incorporated
116 isotopic tracers into such a model, in which evaporative fluxes are assumed proportional
117 to the water vapor and isotopic ratio gradients. Near the water surface within the so-
118 called laminar layer where molecular diffusion is important for vapor transport, the
119 proportionality coefficient, i.e., the diffusivity, differs among different water
120 isotopologues. Above the laminar layer, vapor transport is dominated by turbulent
121 mixing without isotopic fractionation. The model has been widely adopted to quantify
122 isotopic compositions of vapor and/or water in atmosphere, lakes, soil, and plant leaves
123 (see the review by Horita et al., 2008). In using the model, it is assumed that no
124 horizontal vapor gradient exists, so winds have no advective effects on vapor
125 distribution.

126 The no advection assumption is often unrealistic for lakes of limited sizes or
127 other liquid surfaces with limited fetches (e.g., sea ice leads). The presence of advecting
128 air from land and moisture buildup downwind over a lake has been recognized even on
129 rather large water bodies hundreds of kilometers off shore (e.g., Benson and White,
130 1994; Gat et al., 2003). In such cases, wind blowing over the lake picks up moisture from
131 the lake, and as the air advects over the lake towards the downwind shore humidity
132 increases. The moisture content in the air above the lake may exhibit a distribution

133 similar to that illustrated in Figure 1. At a given location on the lake, the moisture
134 content decreases with height, because evaporation occurs at the surface. A horizontal
135 vapor gradient develops as well, with humidity increasing with the distance from the
136 upwind shore as a result of addition of vapor by evaporation (Machavaram and
137 Krishnamurthy, 1995; Gat et al., 2003). In part because wind speed (advection rate)
138 increases with height, the vertical gradient changes with the fetch. Similar to the vapor
139 content, the isotopic composition of vapor in the atmosphere above the lake also has
140 horizontal and vertical gradients. Under such conditions, the spatially averaged vapor
141 and isotopic fluxes from the lake depend on lake size.

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145 Figure 1. Hypothetical vapor concentration distribution above a lake near
146 the upwind shore. Both vertical and horizontal gradients develop (darker
147 shades representing higher vapor concentrations), due to evaporation
148 from the surface and advection of air over the lake by winds that increase
149 in speed with height. The sketch also illustrates one example of our
150 equipment deployment, where the weather station is installed at the
151 upwind shore and vapor measurements using a water vapor isotope
152 analyzer (WVIA) are taken at the downwind (as well as the upwind) shore.

153

154 This project was designed to study the evaporation process for Arctic lakes under
155 conditions shown in Figure 1, where a full quantification of evaporative fluxes requires
156 measurements and modeling in two dimensions. The objectives of this study are to 1)
157 develop a method for measuring horizontal and vertical distributions of the

158 concentration and isotopic ratios of water vapor over lakes, 2) create a physically-based
159 model to explain observed vapor and isotopic distributions, 3) quantify and compare our
160 modeled evaporative fluxes under advection vs. no advection conditions, and compare
161 the latter results with those of a conventional Craig and Gordon type model (1965), and
162 4) understand mechanisms that result in the differences among the model results.

163 **2. METHODS**

164 **2.1 Site Description**

165 The study was conducted in August 2011 near the town of Kangerlussuaq,
166 Greenland located at 67.0212°N, 50.6870°W (Figure 2). The mountainous terrain is
167 mostly tundra covered, and the Greenland ice sheet is approximately 30 km to the east.
168 Kangerlussuaq experiences 15 cm of annual precipitation on average, most falling
169 between June and September. Lakes cover about 20% of the region's surface. Additional
170 water bodies in the landscape include Søndre Strømfjord, (Figure 2), and two rivers fed
171 by melting glaciers.

172 There were several reasons why the experiment was conducted in the Arctic.
173 First, the experiment requires relatively steady wind in both direction and strength.
174 Winds during the summer months in Kangerlussuaq are generally from either the east-
175 northeast from the ice sheet (katabatic winds) or from the southwest along the fjord
176 (sea breezes) (Kopeck et al., 2014). Second, the Arctic has relatively sparse and low
177 stature vegetation, and moisture from transpiration does not cause significant
178 complications in vapor flux measurements. In addition, the low stature shrub and grass
179 dominated vegetation offers relatively steady atmospheric boundary layer conditions.

180 Finally, it was our hope that the method developed in an Arctic area would help
181 understand evaporation from sea ice leads and in cold regions in general.

182 The majority of the study sites investigated is located along the road running
183 from the Kellyville harbor to the ice sheet (Figure 2, grey line). The water bodies studied
184 were lakes 40 m to 6 km long. Average observed wind velocities at 1 m height ranged
185 from 0 to 3.5 m/s. Lake temperatures were typically near 15°C, and air temperature
186 ranged from 12 to 20°C.

187

188 Figure 2. Kangerlussuaq area map. Study lakes, numbered 1-3, are
189 reported in this paper. Roads are shown in grey lines, and the black line
190 south of Lake 3 in the lower panel indicates the section of the road, along
191 which vapor measurements were taken (see 2.2.3).

192

193 2.2 Field Observations

194 At each lake (Figure 2, Table 1), we made meteorological measurements,
195 including temperature, humidity, and wind velocity. Water samples were collected for
196 isotopic analyses. Typically, liquid samples were taken at the shore, but if possible
197 multiple surface samples were also taken off shore. We used a Los Gatos Research
198 Water Vapor Isotope Analyzer (WVIA) to record water vapor mixing ratio (in ppm by
199 volume) and isotopic ratios, relative to V-SMOW, of hydrogen (δD) and oxygen ($\delta^{18}\text{O}$).
200 The calibration of the WVIA has been described by Lauder (2013) and Kopec et al. (2014);
201 the reported 1σ uncertainties of 10 s measurements were 0.44‰ for $\delta^{18}\text{O}$ and 2.2‰ for
202 δD . Deuterium excess (or d-excess), defined as $\delta\text{D} - 8\delta^{18}\text{O}$, is also studied. Given the
203 uncertainties reported for $\delta^{18}\text{O}$ and δD , the 1σ uncertainty for d-excess is 4.2‰ for 10 s
204 measurements.

205 The WVIA samples air through a 1/8" inlet hose. Measurements were taken at a
206 frequency of 0.1 Hz. The delay time of the WVIA reading with varying tube lengths was
207 measured in the laboratory. We found the delay time to be 1 minute, independent of
208 the sampling tube length up to 4 m. In the field, when measurements were taken at
209 different heights, the inlet tube of the WVIA was placed at the desired height for three
210 to twenty minutes, allowing sufficient time for flushing. The transient readings have
211 been removed during post processing.

212 Spatial patterns of vapor measurements included vertical and horizontal lines;
213 for a given lake, the measurements could be of one or both types depending upon
214 logistic conditions. Although vertical vapor distribution measurements at the down wind
215 shore were attempted for ~10 locations, most did not yield reproducible profiles

216 primarily due to variable or changing wind conditions. Here, we report results only for
217 those locations where a stable pattern was obtained through repeated measurements.
218 The location and information about these lakes are shown in Figure 2 and Table 1, and
219 details of the observational method at each lake are described below.

220 *2.2.1 Lake 1*

221 Our most comprehensive observations were made at Lake 1 on two consecutive
222 days. Before the measurement periods, a Davis Vantage Vue weather station was
223 installed on the upwind shore 4.5 m above the lake water surface. The station produced
224 continuous weather records at a frequency of one reading per minute.

225 Measurements over the lake were performed onboard a 2.3 m inflatable raft
226 with an electric motor. Onboard the boat, we used a hand-held weather station to
227 measure humidity, wind velocity, and temperature at 1 m height over the water at
228 multiple locations as we crossed the lake. These instantaneous observations were
229 recorded approximately once per minute. Water temperature was recorded along with
230 the meteorological measurements. Water vapor concentrations and isotopic ratios were
231 obtained for heights alternating between 0.5 and 1.2 m. Vapor profiles, both vertical
232 and horizontal, were reproducible among repeated trips across the lake, indicating
233 steady state conditions.

234 *2.2.2 Lake 2 and the Harbor*

235 At Lake 2 and the harbor, only a vertical profile was recorded on shore. Weather
236 observations were made with the handheld weather station; water temperature was
237 measured and water samples were taken at the shore. For vapor mixing ratio and

isotopic measurements, we attached the sampling tube to a rod and switched between heights at about five-minute intervals. The measurement heights were 0.05, 0.5, 1, and either 2 or 3 m. After sampling from each height starting with the lowest, we repeated measurements at a lower height to verify that the profile was steady over time.

2.2.3 Lake 3 (*Lake Ferguson*)

Lake Ferguson, called Tasersuatsiaq locally, is the largest lake (Lake 3 in Figure 2) near Kangerlussuaq. The Davis Vantage Vue weather station was installed on the downwind shore of the lake. Waves were high, making sampling from the boat unfeasible far from shore. However, the easterly wind made it useful for us to travel by road along the southwest edge of the lake, and sample air that had traveled different distances over water. We carried the WVIA aboard a truck along the road paralleling the southern coast for about 2 km (Figure 2). The observations took about 20 minutes (round trip). Unfortunately, while downwind weather measurement was intensive, we were unable to perform upwind data collection due to logistic constraints.

2.3 Lake Water Isotopic Analyses

The water samples collected at lakes were analyzed at the stable isotope laboratory of Dartmouth College, using a Thermo Finnigan Delta Plus Isotope Ratio Mass Spectrometer (IRMS) interfaced with an HDevice for measuring deuterium to hydrogen ratios and a GasBench for ^{18}O to ^{16}O ratios of water. Results are reported as per mil deviations from the correspondent isotopic ratios of V-SMOW, on the SMOW-SLAP scale, with measurement uncertainties (1σ) of $\pm 0.5\%$ for δD and $\pm 0.1\%$ for $\delta^{18}\text{O}$.

2.4 Model Description

260 We used a two dimensional (2-D) model to simulate vertical as well as horizontal
 261 concentration and isotope distributions of water vapor above a lake. The following
 262 steady state, mass conservation equation is used as the governing equation,

$$263 \quad 0 = \frac{\partial}{\partial z} \left[(D + kz) \frac{\partial C}{\partial z} \right] - u \frac{\partial C}{\partial x} \quad (1)$$

264 where C is the concentration (ppm by volume) of a given isotopologue (i.e. H_2^{16}O , HDO ,
 265 or H_2^{18}O), and D ($\text{m}^2 \text{s}^{-1}$) is the corresponding molecular diffusivity, which is different for
 266 each isotopologue. The variable z (m) is height above the water surface. Borrowing from
 267 Shephard's (1958) generalization of Prandtl (1904), the diffusivity in the vertical
 268 direction is $D + kz$, where D represents the purely molecular diffusion that characterizes
 269 the laminar layer, while kz represents diffusion by turbulent eddies and increases
 270 linearly with height at a rate k . We refer to k ($\text{m}^2 \text{s}^{-1} \text{m}^{-1}$) as the turbulent diffusivity
 271 gradient, with a high k corresponding to a rapid increase of turbulent diffusivity with
 272 height. The variable x (m) is the distance to the upwind shore, and u (m/s) is the wind
 273 speed that, in general, is a function of height, z . Molecular diffusion of water vapor
 274 isotopologues, D , has been established by laboratory experiments, with heavy
 275 isotopologues having lower diffusivities (e.g. for $\text{H}_2^{16}\text{O} \approx \text{H}_2\text{O}$, $D = 2.4 \times 10^{-5} \text{ m}^2 \text{s}^{-1}$ at 15°C ,
 276 $D_{\text{H}_2^{16}\text{O}} / D_{\text{H}_2^{18}\text{O}} = 1.0285$) (Bolz and Tuve 1976; Merlivat 1978). While molecular diffusivity
 277 (D) is different for each isotopologue, the turbulent diffusivity (kz) is not. Below the
 278 transition height $z_t = D/k$, where the molecular and turbulent diffusion coefficients are
 279 equal, molecular diffusion dominates the vertical flux; turbulent diffusion dominates
 280 above z_t . Kinetic isotopic fractionation, which occurs as a result of different D values for

different isotopologues, mostly occurs at heights below z_t . The turbulent diffusion affects isotopic fractionation significantly, but only indirectly by controlling the thickness of the laminar layer and the amount of time vapor remains in that layer. This approach simplifies the solutions to our model, allows comparisons between advective and non-advective scenarios, and does not incorporate a physical discontinuity at the upper boundary of the laminar layer.

Our model assumes that the wind speed at a given height is constant across the lake. As wind carries air from land onto a lake, the decrease in surface roughness between land and water typically causes the wind speed, and possibly turbulent mixing intensity, to increase (See Results). The vertical wind necessitated by conservation of mass would require a vertical advection term in Eqn. 1; this would complicate the model significantly. However, we estimated the vertical advection term to be negligible compared to the included terms. Consequently, we neglect wind speed changes in the x -direction. By making this assumption, we gain advantages of computational simplicity and more accessible physical insights, particularly when we make comparisons between advection and no advection scenarios (section 4).

Solving Eqn. 1 requires knowledge of the wind speed profile, $u(z)$. To obtain $u(z)$, we use conservation of momentum, similar in form as (1), for non-accelerating wind in the x direction,

$$0 = \frac{\partial}{\partial z} \left[(v + k_u z) \frac{\partial(\rho u)}{\partial z} \right] \quad (2)$$

where C in (1) is replaced by momentum density, ρu , the product of air density and wind speed. We took ρ to be constant, as temperature varied little within the air modeled and pressure decreases negligibly in the small height range considered. The molecular diffusivity for momentum density is the kinematic viscosity of air (ν), which is a known temperature-dependent quantity (e.g. $\nu = 1.5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ at 15°C , Montgomery, 1947). The parameter k_u ($\text{m}^2 \text{ s}^{-1} \text{ m}^{-1}$) is the equivalent of k in (1), the gradient of momentum diffusivity. When wind speed at height (0) and another height (z_1) are known we can express the boundary conditions as $u(0) = u_0$, and $u(z_1) = u_1$, and (2) has the following closed form solution,

Figure 3. The vertical wind profile from Eqn. 3, using wind observations at Lake 1, Day 1.

$$u(z) = \frac{u_1 \ln\left(\frac{\nu}{\nu + k_u z}\right) + u_0 \ln\left(\frac{\nu + k_u z}{\nu + k_u z_1}\right)}{\ln\left(\frac{\nu}{\nu + k_u z_1}\right)} \quad (3)$$

If wind speeds at two heights above the surface are known, and with the assumption $u_0 = 0$, we can also solve for k_u . An example solution for $u(z)$ based on wind speeds

318 recorded at Lake 1 (Day 1) is shown in Figure 3. Such a profile was used in Eqn. 1 for
319 solving vapor concentration and isotopic ratios.

320 To solve for water vapor concentration across a lake, we use a lower boundary
321 condition that assumes that the water vapor concentration at $z = 0$ is at saturation (C_s)
322 (ppm by volume). We obtained the saturation value, C_s , by converting the saturation
323 specific humidity (q_s , mg/g), expressed as a function of absolute temperature (T) by
324 Glanz and Orlob (1973) as cited by Condie and Webster (1997),

$$325 \quad q_s = 1.35 \times 10^5 \exp\left(\frac{-4157}{T - 33.91}\right) \quad (4)$$

326 For the boundary condition at the upwind shore, we used the average
327 concentration and isotopic ratios at 1 m (1.2 for Lake 1) as the representative values
328 that distribute homogeneously with height up to the top (10 m) of the model. This
329 assumption is not strictly valid, particularly near the surface, because concentrations of
330 isotopologues, like wind, likely have a vertical profile at $x=0$. This can be seen from the
331 concentration and isotopic differences measured at (or near) the shore between 0.5 and
332 1.2 m height (Figure 5). The difference, however, is not large, particularly not generally
333 larger than the scatter of the data. One exception is d-excess, which does change
334 significantly with height at the shore. In this model we ignore this difference for
335 simplicity. Our assumption should be adequate for $z=1$ m and higher, because most
336 vertical change is likely limited to near the surface. The top is set to 10 m. This choice is
337 sufficiently high that the solution is not sensitive to this condition (in our simulations,

338 changing the top to 5 or 15 m does not affect the solution) yet sufficiently low that the
 339 computational cost of the model is not impractically large.

340
 341 Figure 4. A sample solution of Eqn. 1: isopleths of total water vapor
 342 concentration (ppm), using boundary conditions for Lake 1 (Day 1).
 343 Concentrations are modeled above the water surface over a 500 m fetch for
 344 $k=7.6 \times 10^{-3} \text{ m}^2 \text{ s}^{-1} \text{ m}^{-1}$. Concentrations for isotopologues of H_2^{18}O and HDO can be
 345 similarly modeled using the corresponding molecular diffusivity values.

346
 347 Using known molecular diffusion coefficients (Bolz and Tuve 1976; Merlivat
 348 1978), 2-D concentration profiles can be calculated (example shown in Figure 4). In
 349 simulating a given data set, k is selected to create the best match between computed
 350 results and the observational data by the least square method. Numerical solutions to
 351 our governing equation were calculated using the method of lines in Mathematica®
 352 computational software (see Appendix A for supplementary material). Some parameters,
 353 such as grid size (and height of the top previously mentioned), are adjusted depending
 354 upon the conditions and scale of the site being modeled to achieve minimum sensitivity
 355 of the solution to the parameters.

356 Once Eqn. 1 is solved, the evaporation rate and evaporative isotope fluxes can be
357 determined. For the bulk vapor, the vertical flux $F(z)$ is

$$358 \qquad F(z) = -(D + kz) \frac{\partial C}{\partial z} \qquad (5)$$

359 At $z = 0$, this flux becomes the evaporation rate, E , from the water surface.

$$360 \qquad E = F(0) = -D \frac{\partial C}{\partial z} \bigg|_{z=0} \qquad (6)$$

361 The evaporative flux can also be determined for individual isotopologues using Eqns. 5
362 and 6, and thus flux ratios expressed in δD and $\delta^{18}O$ units can be calculated.

363 364 **3. RESULTS**

365 **3.1 Observational Data**

366 Several general observations apply to all the lakes we observed. Air upwind of all
367 the lakes was dry and depleted in deuterium and ^{18}O relative to vapor in equilibrium
368 with the lake water. Therefore, during this study, humidity and isotopic ratios over all
369 lakes were generally highest near the water surface and near the downwind shore. As
370 the air moves across the lake, gaining moisture from the surface, it increases in both
371 humidity and isotopic ratios. Thus, the combination of evaporation, vertical mixing, and
372 horizontal advection causes a humidity increase in both the downwind and vertical
373 downward directions associated with isotopic ratio increases in the same directions.

374 The water and air measurements are summarized in Tables 1a and 1b. When we
375 were able to measure surface water temperature at the center of the lakes, we found it
376 to be the same as that at the shore. Similarly, there was no significant difference in δD

and $\delta^{18}\text{O}$ values of water between near shore and mid lake samples. Compared to the moisture concentration, isotopic ratios in the air seemed more sensitive to small changes in wind direction. The sampling periods that yielded reproducible results were those of relatively steady wind with regular gusts such that sustained 10-s wind speed did not exceed two times the average wind speed over the observation interval. Detailed observations at each lake are described below.

Table 1a. General Observations of Study Lakes

Site	Fetch Length (m)	Water δD (‰)	Water $\delta^{18}\text{O}$ (‰)	Water Temp. ($^{\circ}\text{C}$)	Air Temp. ($^{\circ}\text{C}$)	Wind Speed (m/s) ⁽¹⁾	Observation ⁽²⁾ Method
Lake 1 (Day1)	500	-94.9	-9.0	15	16	3.2	H&V on water
Lake 1 (Day2)	500	-94.9	-9.0	15	12	2.2	H&V on water
Lake 2	150	-88.3	-6.6	14.8	19.5	1.2	V on shore
Lake 3	5600	-111.3	-15.0	11.7	19	5	H on shore
Harbor	N/A	-99.0	-8.4	8.2	12	0.6	V on shore

(1) Average 1 m wind speed, except for Lake 3 which was at 5 m height.

(2) H=horizontal, V=vertical.

Table 1b. Upwind Air Measurements

Site	H ₂ O (ppm)	H ₂ O stdev	δD ‰	δD stdev	$\delta^{18}\text{O}$ ‰	$\delta^{18}\text{O}$ stdev
Lake 1 (day 1)	5727	268	-234.5	2.0	-34.2	0.5
Lake 1 (day 2)	5887	253	-242.0	3.3	-35.7	1.4
Lake 2	5148	102	-237.3	1.7	-33.7	0.4

3.1.1 Results from Lake 1

The data from both observation days for Lake 1 are shown in Figures 5. We experienced greater wind speeds at the center of the lake than at the shore. The on-shore weather station at a 4.5-m height recorded data continuously on day 1; the average wind speed was 4.5 m/s, with gusts up to 6.5 m/s. The frequency of gusts and

395 average wind speed varied little. On both days, we observed expected increases in
396 humidity and isotopic ratios with increasing distance from the upwind shore and
397 decreasing height above the water surface (Figure 5). Over the fetch of the lake at 1.2 m
398 height, the vapor content increased with distance at a rate of about 1380 ppm/km, δD
399 at 21‰/km, and $\delta^{18}O$ at 2.4‰/km. More fluctuations occurred at 0.5 than 1.2 m height.
400 Deuterium excess (Figure 5) shows different patterns for measurements at 0.5 vs.
401 1.2 m. On Day 1, d-excess at 1.2 m decreased significantly ($p < 0.0001$) at a rate of
402 5‰/km, while at 0.5 m, it increased more or less linearly at about 24‰/km. At
403 distances greater than 300 m, the d-excess of the two heights overlapped. Data on Day
404 2 show similar patterns with two differences: 1) the decreasing trend for 1.2 m data was
405 the same 5‰/km, but because of the smaller quantity and greater variability of data the
406 trend is not statistically significant ($p = 0.16$), and 2) the 0.5 m data show very rapid
407 increase in d-excess at short distances (< 50 m), merging with, although still below, the
408 1.2 m data at distances greater than 200-300 m.
409

410
 411 Figure 5. Lake 1 vapor profiles obtained on Day 1 (left) and Day 2 (right). Air at
 412 0.5 m above the water surface (triangles) was generally more concentrated in
 413 all three isotopologues than that at 1.2 m (circles). Humidity, δD and $\delta^{18}O$ all
 414 increased as air traveled downwind over water. Deuterium excess (lowest
 415 panels), however, show different trends with distance at 0.5 m (increasing) vs.
 416 1.2 m (decreasing) heights.

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Figure 6. Dependence of 1.2 m vapor concentration and isotopic ratios on the mean wind speed. The vertical axis is the residual of vapor concentration or isotopic ratios with their linear dependency on the distance off shore removed. The horizontal axis is the average 1-minute wind speed at Lake 1 (Day 1).

425 There was a tendency for the turbulent transport of moisture to increase when
426 wind intensified. When H₂O concentration is regressed against the distance from the
427 upwind shore, the height of measurement and the wind speed, the overall multiple
428 regression explains 78% ($p < 0.0001$) of the variance after two outliers are excluded. Each
429 of the three explanatory variables independently explains a statistically significant
430 fraction of the H₂O variance. The same results were also obtained for either $\delta^{18}\text{O}$
431 ($r^2 = 0.85$) or δD ($r^2 = 0.83$) in place of the H₂O concentration. Figure 6 shows the linear fit
432 between humidity and isotopic ratios residuals at 1.2 m height (after their linear
433 dependency on distance from shore was removed), and the 1-minute average wind
434 speed (measured on shore). The regression is significant for vapor concentration and δD
435 ($p < 0.05$), marginally significant for $\delta^{18}\text{O}$ ($p < 0.10$), while not significant for d-excess
436 ($p = 0.6$). We note that vapor isotopes were measured at various locations over the water
437 surface, while wind speed was measured only on the upwind shore. Correlating these
438 two quantities recorded at locations separated by 50 to 500 m might have contributed
439 to part of the scatter in the figure.

440 *3.1.2 Results from Lake 2 and the Harbor*

441 Figure 7 contains results for Lake 2, where moisture content and its isotopic
442 composition at four heights on the downwind lakeshore clearly show a decrease with
443 increasing height within the first two meters from the surface. The d-excess shows an
444 opposite trend, increasing with height. The total ranges of the change with height are
445 about 4,000 ppm, 35‰, 6‰, and 7‰ for vapor concentration, δD , $\delta^{18}\text{O}$, and d-excess,
446 respectively, with greater gradients near the surface.

447

448 Figure 7. Vertical profiles of vapor quantities for Lake 2. Values with standard
 449 deviations were recorded at the downwind shore. Measurements were
 450 repeated at 0.5 m, indicated by two sets of points with correspondent error
 451 bars at that height. Deuterium excess was computed from δD and $\delta^{18}O$ data
 452 with propagated error bars. Model fits are shown in solid curves (Section 3.2).
 453 The observations at the lowest height (0.05 m) were not used for obtaining the
 454 best-fit k value. Their inclusion did not change the estimated k value, but
 455 significantly increased the error of the estimate.

456

457 The measurement method at the Harbor site was the same as that at Lake 2. This
 458 site (and one of the other lakes not reported here) experienced low to no wind and high
 459 humidity conditions, which resulted in low vapor and isotopic gradients. While these
 460 vapor profiles were consistent with our expectations, calculations of the evaporative
 461 fluxes were associated with significant uncertainties. Therefore, we will not consider the
 462 Harbor site further in the discussion.

463 3.1.3 Results from Lake 3

464 At this lake, wind conditions varied over the three-hour period of meteorological
 465 observation; however, there were several thirty-minute intervals of relatively steady
 466 conditions. Changes in water vapor concentration and isotopic ratios showed small but
 467 statistically significant decreases as we traveled upwind (Figure 8); the horizontal

468 gradient is 354 ppm/km ($p < 0.0001$) for vapor content, 1.5‰/km ($p = 0.0032$) for δD ,
469 0.22‰/km ($p = 0.013$) for $\delta^{18}O$, and 0.3‰/km ($p = 0.28$) for d-excess. These gradients are
470 all smaller than the corresponding gradients at Lake 1, where the fetch was much
471 smaller. The r^2 values (see Figure 8), however, are small, indicating that the influence of
472 distance upwind on vapor content and isotopic ratios is small relative to other random
473 effects. Although the road varied in elevation from 7 m to 30 m, we did not observe a
474 significant vertical change in the vapor composition.

475 *3.1.4 Summary of Observations*

476 Generally, the water vapor concentration and isotopic ratios in air all increased
477 as the air traveled across a lake downwind, and they were highest near the water
478 surface, decreasing with height. This trend was observed at fetches ranging from 40 m
479 to over 5 km. Deuterium excess generally exhibited a trend opposite to those of δD and
480 $\delta^{18}O$. However, at Lake 1, d-excess at 0.5 m increased with distance while that at 1.2 m
481 decreased. We also observed a significant positive partial correlation between wind
482 speed and concentrations of all three vapor isotopologues at Lake 1.

483

484

485 Figure 8. Observations of vapor and isotopic ratios at Lake Ferguson. The data
 486 were collected along the southwest side of the lake as shown in Figure 2. With
 487 increasing distance downwind, water vapor content and isotopic ratios all
 488 significantly increased, while d-excess decreased (although not statistically
 489 significant). The linear regression line is added to each of the plots.

490 3.2 Model Results

491 The modeling effort involved two steps. In the first step, we obtained the best-fit
492 k value by fitting the observational data to solutions of Eqn. 1 for each given lake. We
493 then use the solution with the best-estimate k to calculate vapor and isotopic fluxes that
494 are consistent with our observed concentration and isotopic ratio distributions. Because
495 of the inhomogeneous boundary conditions at the lakeshore, as discussed earlier, we
496 simulated observations of Lake 1 only at 1.2 m, where our assumptions are likely valid.

497 We calculated the sum of squares of differences between observed and modeled
498 concentrations of isotopologues to yield best-fit k using the method of least squares.
499 With three vapor isotopologues, rather than one, we achieved stronger constraints on
500 the value of k . Figure 9 shows that the sum of squares of errors (SSE) of the
501 concentration of each isotopologue as a function of k , and the three SSE's combined.
502 The latter provides a more accurate estimate of the best-fit value of k , indicated by
503 steeper slopes of the SSE curve around the better-defined minimum. Combining the
504 three vapor and isotopic observations led to approximate halving of the uncertainty in k
505 for each of the observation periods due to increased number of data points, the greater
506 sensitivity of the isotopes to k , and the precision of isotopic data.

507

508 Figure 9. Normalized sum of squares error (SSE) of model fit to data vs. k for Lake
 509 1 (Day 2) (upper) and Lake 2 (lower). The SSE for each quantity is normalized by
 510 dividing by the data variance. The best-fit k values for Lake 1 Day 1 (not shown
 511 here), Lake 1 Day 2 and Lake 2 are $7.6 \pm 0.5 \times 10^{-3}$, $4.9 \pm 0.5 \times 10^{-3}$, and $2.8 \pm 0.1 \times 10^{-3}$
 512 $\text{m}^2 \text{s}^{-1} \text{m}^{-1}$, respectively.

Figure 10. Model fits to horizontal vapor profiles at 1.2 m height based on data of Lake 1 on Day 2 with $k=4.9 \times 10^{-3} \text{ m}^2 \text{ s}^{-1} \text{ m}^{-1}$.

518 The best-estimate k values were used to generate the vapor profiles.

519 Comparison between the data and model simulations are shown in Figures 7 and 10.

520 Using Eqn. 5, the distribution of fluxes (Figure 11) was computed using the solutions of

521 Eqn. 1 with the best-fit k . The evaporative vapor flux ($z=0$) decreases with distance

522 downwind, and isotopic ratios increase. Note that the decrease along the fetch for δD

523 (4‰) is comparable (and slightly smaller) in magnitude to that of $\delta^{18}O$.

524

525 Figure 11. Modeled vapor and isotopic ratios of vertical fluxes above Lake 1 on Day 2.

526 The Isopleths of fluxes of H_2O are in cm/year, and isotopic ratios of fluxes are in per mil

527 (‰).

528

529 4. DISCUSSION

530 There are several indications verifying the adequacy of the 2-D model's

531 representation of physical processes of lake evaporation that we observed. First, the

532 model produced both vertical and horizontal moisture and isotopic gradients consistent

533 with our observations; the vapor concentration and isotopic ratios increased with

534 distance from the upwind shore (Figure 10) and decreased with height (Figure 7).

535 Second, the diffusivity gradient k yielded from our model simulations are well
536 within the range reported in previously work. The least-squares estimates of turbulent
537 diffusivity gradient, k , using our model convert to 1 m turbulent Austausch coefficients
538 of 2.8×10^{-3} , 4.9×10^{-3} , and $7.6 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$, while a range from 5×10^{-4} to $1 \text{ m}^2 \text{ s}^{-1}$ was
539 reported by Richardson (1921). Furthermore, our three estimates of 1 m Austausch
540 coefficients, corresponding to wind speeds of 1.2, 2.2 and 3.2 m/s, respectively (Table
541 1a), have a positive ordinal correlation with the 1 m wind speed, as expected by both
542 boundary layer theory and experiments (e.g., Sutton, 1932). A similar result has been
543 observed in a recent study of Arctic lakes (Granger and Hedstrom 2011).

544 Third, our model produced reasonable evaporation rates. The vertical flux of
545 moisture for Lake 1 Day 2 is illustrated in Figure 11 (panel H_2O). The evaporation rate is
546 found by calculating the vertical flux at height zero. The values for all lakes range from
547 20 to 60 cm/year (annualized) or 0.5 to 1.6 mm/day. These rates are comparable, within
548 error, to the observed fluxes from Arctic lakes (Granger and Hedstrom 2011).

549 Fourth, the model explains well the d-excess observed at 1.2 meters. The upwind
550 air is dry and high in d-excess. For example, on Day 1 at Lake 1, the d-excess of upwind
551 air at the shore is $\sim 39\%$ (Table 1b). This value decreases gradually over the lake as
552 evaporative moisture with lower d-excess is mixed into the dry air. However, with the
553 simple assumption that the upwind air is vertically homogeneous, the 2-D model does
554 not explain the horizontal d-excess profile at 0.5 m for Lake 1, which showed an increase
555 with distance rather than a decrease expected by model. This is probably because the d-
556 excess at the boundary exhibits a vertically increasing trend, which is particularly

557 prominent near the surface. However, the estimate of the best k values is not sensitive
558 to this assumption. We suggest that the vertical distributions of vapor content and
559 isotopic ratios be investigated in greater detail with more levels of measurements in
560 future work. A more accurate boundary condition can then be applied. With this model
561 verified by observations, we assess below the effect of advection on the evaporative
562 isotopic fluxes, and quantify potential biases of doing lake water mass balance
563 calculations that require isotopic ratios of evaporative flux.

564 4.1 Effect of Advection on Isotopic Ratios of Evaporative Vapor Flux

565 Using the model, we can compare evaporative fluxes with and without advection.
566 Advection, both horizon and vertical, is not typically considered in evaporation models
567 incorporating water isotopologues. Both the original Craig and Gordon (1965) model
568 and the more frequently used Merlivat and Jouzel expression (1979) are based on
569 horizontally uniform steady-state conditions. However, in our study over small bodies of
570 water, humidity was not horizontally uniform, which is demonstrated by the
571 observations as well as the two dimensional model results. Recall that we do not
572 consider vertical advection, although a more general treatment of the planetary
573 boundary layer, has been developed and will be published separately. To compare the
574 isotopic fluxes from the 2-D model with those where advection is absent, we can simply
575 set the wind velocity in (1) to zero, and the equation reduces to one-dimensional
576 (vertical). This 1-D equation, in combination with the boundary conditions $C(0) = C_s$

577 (surface saturation) and $C(z_1) = C_1$ (a known or specified value), has the closed form
 578 solution,

$$579 \quad C(z) = \frac{C_1 \ln\left(\frac{D}{D+kz}\right) + C_s \ln\left(\frac{D+kz}{D+kz_1}\right)}{\ln\left(\frac{D}{D+kz_1}\right)} \quad (7)$$

580 Taking the derivative of (7) with respect to z , the vertical concentration gradient is

$$581 \quad \frac{\partial C}{\partial z} = \frac{k(-C_1 + C_s)}{(D+kz)(\ln D - \ln(D+kz_1))} \quad (8)$$

582 Eqn. 8 can be substituted into Eqn. 5 to determine the flux of each of the three
 583 isotopologues at any height (although in the 1-D case fluxes are independent of height),
 584 and the evaporative fluxes are obtained by substituting $z = 0$. The δD and $\delta^{18}O$ values of
 585 the evaporative flux can then be obtained using the δ definition. Our test scenario to
 586 compare the 1-D and 2-D models was based on conditions observed at Lake 1 on Day 2.
 587 The upwind vapor quantities (Table 1) were assigned to $z_1 = 10$ m to be comparable to
 588 the 2-D situation. The results of these model runs are listed in Table 2. The range of
 589 surface fluxes summarized in the last column of Table 2 comes from the same advection
 590 scenario shown in Figure 11.

591 Table 2 shows that the advection scenario is associated with significantly more
 592 negative isotope flux values compared to no advection calculations. For this particular
 593 example, $\delta^{18}O$ and δD are both lowered by up to $\sim 4\text{‰}$ when advection is incorporated.
 594 Clearly, this isotopic depletion is a result of the influence of the dry air from the land,
 595 causing intensive kinetic fractionation. Such an isotopic contrast between advection and

no advection conditions should decrease with increasing fetch (Figure 11) as the land vapor continuously equilibrates with the lake water downwind. Asymptotically, the 2-D values should approach the 1-D values. At Lake 3, the isotopic gradient within 4-5 km from the upwind shore was 354 ppm/km, 1.5‰/km and 0.22‰/km, for the vapor concentration, δD and $\delta^{18}O$, respectively. These are smaller than the corresponding gradients observed at Lake 1 within 0.5 km from the upwind shore - 1380 ppm/km, 21‰/km, and 2.4‰/km, respectively. Wind speed is another factor that influences the maximum fetch over which advection causes changes in evaporative fluxes. However, this effect is difficult to quantify because, the outcome depends upon an integrated effect of wind speed on the magnitude of k , on the vapor gradient, and on the rate of vapor-water equilibration, all of which may not have the same sign. Unfortunately, we do not have the upwind data for Lake 3 to allow a model simulation under a higher wind speed and a longer fetch.

Using the 1-D no advection model, we can also compare the isotopic fluxes with those yielded by 0-D models. We use the Merlivat and Jouzel (1979) model result (their Eqn. 9) for this comparison. The input parameters required by the Merlivat-Jouzel model are the wind speed at a specific height, relative humidity, isotopic compositions of the air and water, and water surface temperature. However, being 0-D, the Merlivat-Jouzel model does not specify the height at which the relative humidity and the isotopic ratios should be observed. We used our upwind measurements for these quantities (Table 1). For the 1-D model, we used $4.9 \times 10^{-3} \text{ m}^2 \text{ s}^{-1} \text{ m}^{-1}$ as the turbulent diffusivity gradient (k) based on the result of our advection model simulation for Lake 1, Day 2. Unlike the

Merlivat-Jouzel model, our model does not neglect changes of humidity with height. In Table 2 we show results for two no advection calculations for which the 34.4% relative humidity (RH) and isotopic ratios of upwind air was set to be at 0.2 and 10 meters, respectively.

Table 2. Comparison of Evaporative Isotopic Fluxes among 0-D and 1-D (No Advection) and 2-D (Advection) Scenarios

Variables	Merlivat and Jouzel (1979)	This Model ($k=4.9 \times 10^{-3} \text{ m}^2 \text{ s}^{-1} \text{ m}^{-1}$)		
		No advection with 0.344% RH (at z_1 for this model)		Advection
z_1 (m)	n/a	0.2	10	n/a
Flux δD (‰)	-138.8	-138.8	-136.0	-137 to -140
Flux $\delta^{18}\text{O}$ (‰)	-17.5	-17.5	-14.0	-15 to -18

Our 1-D flux values are identical to those of the Merlivat-Jouzel model only when z_1 is set at one fifth of a meter (0.2 m), while higher z_1 yielded less negative δ values. The magnitude of enrichment was particularly significant for $\delta^{18}\text{O}$ with up to a 3.5‰ increase for z_1 changing from 0.2 to 10 m. In the 1-D model, when z_1 is set higher than 0.2 m, the 0.2 m relative humidity (nearer the surface) would be greater than 34.4%, causing isotopic fluxes to be more enriched in both deuterium and ^{18}O , the latter being more sensitive to this effect due to its more pronounced kinetic fractionation (Merlivat, 1978). Therefore, if there were a humidity gradient above the water surface, the humidity measurement at different levels would yield different isotopic fluxes using the Merlivat and Jouzel model; the actual fluxes are not necessarily the same as the computed values.

Figure 12. Plot of δD vs. $\delta^{18}O$ for all observed and computed components of water and water vapor for Lake 1 Day 2. The inserted smaller graph shows the vapor data in a better resolution. Flux values are from Table 2. The local precipitation (Kopec et al., 2014) is also included. It is on the local meteoric water line, which is different from the global meteoric water line (GMWL) in the graph.

The δD vs. $\delta^{18}O$ relationship for all observed and computed entities are displayed

in Figure 12. We make several comments about this graph.

- 1) The diagram shows the relative enrichment among different components contributing to the isotopic systematics of the water and water vapor system. The lake water has the highest δD and $\delta^{18}O$ values, and the upwind air has the lowest. The second most enriched member is the evaporative fluxes. To first order, mixing of the evaporated vapor and the upwind (input) air controls the observed vapor isotopic ratios.
- 2) The graph also shows the relative differences in d-excess among the components of the system, which is indicated by their displacement above or below the global meteoric

653 water line (GMWL). In this dimension, lake water has the lowest (most negative) d-
654 excess, the upwind air has the highest (most positive), and vapor above the lake and
655 evaporative fluxes are in between. The d-excess of evaporative fluxes ranges from
656 relatively high values near the upwind shore to lower values with increasing fetch. This
657 makes sense because the kinetic isotopic fractionation is expected to be highest near
658 the shore, where the relative humidity is low.

659 3) The slope of the vapor isotope trends is less than 8 (overall slope= 4.4 ± 0.2), consistent
660 with our general observation that d-excess decreases with increasing enrichment of
661 deuterium and ^{18}O in water vapor vertically downward and/or downwind.

662 4) The local precipitation is more depleted in deuterium and likely also ^{18}O than the
663 evaporated water (1-D or 2-D). This is somewhat unusual, and suggests that the input
664 water to the lake must be isotopically more enriched than the local precipitation. One
665 possibility is that some input water is derived from hill slope runoff that has been
666 partially evaporated.

667 668 4.2 Effect of Lake Size on Isotopically Inferred Water Balance

669 When horizontal advection acts on the vapor distribution, lake size influences
670 the relationship between isotopic fluxes and lake water balance. If wind advection is
671 neglected in the inference of water balance from lake isotopes, an error may thus be
672 introduced, the magnitude of which depends on lake size. We refer to this as the “lake
673 size effect”. As a result of the lake size effect, the treatment of advection and the choice
674 of 1-D vs. 2-D models have implications for the inference of lake water mass balance
675 and paleoclimate. There have been significant efforts in modeling variations of lake

676 water isotopic compositions, with a purpose for both hydrological studies (e.g., Gibson
677 et al., 1993; Yi et al., 2007; Turner et al., 2014; Vallet-Coulomb et al., 2006), and
678 paleoclimate reconstructions using the lake isotope records (e.g., Steinman et al., 2010;
679 Jones and Imbers, 2010). In all these studies, the evaporation model is of the Craig and
680 Gordon zero dimensional type. Although it has been acknowledged that advective air
681 would potentially affect isotopic evaporative fluxes for small lakes (e.g., Steinman and
682 Abbott, 2013), the effect has not been assessed quantitatively. In the specific
683 environmental setting of Kangerlussuaq, Greenland, this effect is significant. Figure 13
684 illustrates how the averaged isotopic ratios of the evaporative flux over a given size of a
685 lake differ from the isotopic values of the flux under 1-D (no advection) conditions. The
686 averaged values of δD and $\delta^{18}O$ have been subtracted by the corresponding 1-D values
687 (Table 2), and thus the graph demonstrates additional depletions of the fluxes due to
688 advection relative to the 1-D values. In this example (Lake 1, Day 2) the δD and $\delta^{18}O$ of
689 vapor fluxes, averaged over 500 m along the wind direction, are about 2‰ more
690 depleted relative to the 1-D scenarios. The differences are greater for smaller fetches.
691 Note that the differences between 2-D and 1-D fluxes are similar for δD and $\delta^{18}O$, with
692 the latter being slight greater. This is because the size of kinetic fractionation for oxygen
693 and hydrogen are comparable. This result is important for assessing the bias caused by
694 ignoring advection in water balance calculations.

695 The information in Figure 13 can be translated into potential bias or
696 uncertainties in calculations of water balance of lakes and catchments. In a simplest
697 case with steady state of lake water volume, the ratio of input (I) to evaporation (E)

698 fluxes of the lake can be obtained by the isotopic compositions of the evaporative flux
 699 (δ_E), lake water (δ_L) and local precipitation (δ_P) by (e.g., Gibson et al., 1993)

$$700 \quad \frac{I}{E} = \frac{\delta_E - \delta_L}{\delta_P - \delta_L} \quad (9)$$

701 If we assume that the difference between δ_E and δ_L is roughly the equilibrium difference
 702 between liquid and vapor, then every per mil depletion due to ignored advection would
 703 result in an about 10% underestimation of the I/E ratio if using $\delta^{18}\text{O}$ in Eqn. 9, but the
 704 error is only a little over 1% if using δD . This asymmetry suggests a greater robustness of
 705 using δD in water balance studies and paleoclimate reconstructions, where the lake size
 706 is small and/or changing significantly through the study period.

707
 708 Figure 13. Comparison of 1-D vs. 2-D isotopic fluxes over an entire lake of
 709 different sizes. The 1-D values of both δD and $\delta^{18}\text{O}$ are set to 0‰. The 2-D
 710 values are averaged over the lake, and then the 1-D values are subtracted
 711 from the averages. This result only applies to lakes of idealized rectangular
 712 shape experiencing steady wind parallel to one of the edges. The lake size
 713 plotted here is the length scale long the wind direction. The calculation is
 714 based on the surface fluxes shown in Figure 11.

716 5. CONCLUSION

717 We report concentration and isotopic distributions of water vapor in air over
718 several Arctic lakes near Kangerlussuaq, west Greenland. This investigation featured
719 new methods of data collection using the mobile deployment of a Water Vapor Isotope
720 Analyzer (WVIA). We expected, for lakes of limited fetch, a non-uniform distribution of
721 moisture and isotopic ratios above the lake, with vapor concentration increasing with
722 distance from the upwind shore and decreasing with increasing height. This pattern was
723 systematically observed for lakes that experienced steady winds during the observation
724 period; within the first 0-0.5 km fetch, we observed horizontal gradients of 1380
725 ppm/km for water vapor content, 21‰/km for δD , 2.4‰/km for $\delta^{18}O$, and 5‰/km for
726 d-excess. These gradients decrease with increasing distance from the upwind shore. At a
727 point over 4-5 km from the upwind shore on a different lake, we observed 354 ppm/km,
728 1.5‰/km, 0.22‰/km and 0.3‰/km, for vapor concentration, δD , $\delta^{18}O$, and d-excess,
729 respectively. Vertical moisture and isotopic gradients are seen from measurements at
730 different heights over a lake or at the downwind shore. The size of the vertical gradient
731 depends on wind speed, fetch, and turbulence intensity, being greatest near the surface,
732 decreasing with height.

733 We used a 2-dimensional (2-D) steady state advection diffusion model to
734 simulate the observed distributions of isotopologues, and to calculate their evaporative
735 fluxes. The validity of the model is demonstrated by 1) successful simulation of the
736 observed data, 2) reasonable turbulent diffusivity gradients constrained by fitting the
737 model to the data, and 3) flux of evaporation yielded by model solutions being
738 comparable to similar studies.

739 When horizontal advection velocity is zero, the 2-D model reduces to a 1-D
740 model. Comparing results of these two models, we evaluated the effect of advection
741 and lake size on isotopic ratios of vapor flux from lake evaporation. The result shows
742 that advection, particularly advection of dry air from the land onto the lake, significantly
743 affects the vapor and isotopic fluxes from the lake; our calculations for a particular
744 example yielded up to a 4‰ difference, in both δD and $\delta^{18}O$ values within 500 m fetch,
745 compared to the 1-D (no advection) flux. When averaged over a lake along the wind
746 direction, the vapor flux from the entire lake is more depleted in D and ^{18}O compared to
747 the 1-D flux, with the magnitude of depletion greater for smaller lake size. An example
748 calculation yielded a 2‰ or more difference between 1-D and 2-D scenarios with a fetch
749 up to 500 m. We assessed that, with each 1‰ discrepancy, the lake size effect leads to
750 an underestimation of the I/E (input to evaporation) flux ratio by about 10% using $\delta^{18}O$
751 and ~1% using δD . This argues for advantageous use of δD as climate proxies if the study
752 lake is small or changing in size through its history.

753 The 1-D model can be used to compare isotopic fluxes with those from a classical
754 0-D steady state model commonly used for isotopic flux calculations. By setting the
755 observed relative humidity value of 34.4% at 0.2 m, we were able to produce δD and
756 $\delta^{18}O$ values of vapor flux identical to those yielded by the Merlivat and Jouzel (1979)
757 model. However, by setting the level of the 34.4% relative humidity value at greater
758 heights (e.g., 10 m which is consistent to our 2-D model), we obtained less negative
759 δ values for isotopic fluxes. Since our 1-D model has a vertical moisture gradient, the
760 relative humidity decreases with height. Setting the same relative humidity value at a

761 higher altitude is equivalent to setting a higher relative humidity near the surface, and
762 thus kinetic fractionation is reduced. We conclude that, if using relative humidity values
763 measured at an arbitrary height in the 0-D two-layer steady state model, one may not
764 obtain flux values representative of what is actually occurring in the field. Typically, if
765 using the common meteorological measurements of 2 or 10 m, the calculated isotopic
766 ratios of evaporative fluxes from the 0-D model would likely be too low.

767 The observation methodology and model we developed in this study may be
768 useful for similar investigations over open water under different environmental settings.
769 The impact and sensitivity of advection on evaporative fluxes is likely to vary with
770 relationships among prevailing meteorological and hydrological variables. We showed
771 that, by simultaneous measurements of vapor and its isotopic ratios, we reduced the
772 uncertainty in determining transport coefficients. The results may also help understand
773 evaporation from other water surfaces with limited fetch, such as from sea ice leads.

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775
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783 **Appendix A. Supplementary Material**

784

785 Supplementary material related to this article can be found at:

786

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