Delineating extent and magnitude of river flooding to lakes across a northern delta using water isotope tracers

Casey R. Remmer | Tanner Owca | Laura Neary | Johan A. Wiklund | Mitchell Kay | Brent B. Wolfe | Roland I. Hall

1Department of Biology, University of Waterloo, Waterloo, Ontario, Canada
2Department of Geography and Environmental Studies, Wilfrid Laurier University, Waterloo, Ontario, Canada

Correspondence
Casey Remmer, Department of Biology, University of Waterloo, 200 University Avenue West, Waterloo ON N2L 3G1 Canada.
Email: crremmer@uwaterloo.ca

Abstract
Hydrological monitoring in complex, dynamic northern floodplain landscapes is challenging, but increasingly important as a consequence of multiple stressors. The Peace-Athabasca Delta in northern Alberta, Canada, is a Ramsar Wetland of International Importance reliant on episodic river ice-jam flood events to recharge abundant perched lakes and wetlands. Improved and systematic monitoring of landscape-scale hydrological connectivity among freshwater ecosystems (rivers, channels, wetlands, and lakes) is needed to guide stewardship decisions in the face of climate change and upstream industrial development. Here, we use water isotope compositions, supplemented by measurements of specific conductivity and field observations, from 68 lakes and 9 river sites in May 2018 to delineate the extent and magnitude of spring ice-jam induced flooding along the Peace and Athabasca rivers. Lake-specific estimates of input water isotope composition (δ) were modelled after accounting for influence of evaporative isotopic enrichment. Then, using the distinct isotopic signature of input water sources, we develop a set of binary mixing models and estimate the proportion of input to flooded lakes attributable to river floodwater and precipitation (snow or rain). This approach allowed identification of areas and magnitude of flooding that were not captured by other methods, including direct observations from flyovers, and to demarcate flow pathways in the delta. We demonstrate water isotope tracers as an efficient and effective monitoring tool for delineating spatial extent and magnitude of an important hydrological process and elucidating connectivity in the Peace-Athabasca Delta, an approach that can be readily adopted at other floodplain landscapes.

KEYWORDS
floodplain lakes, freshwater ecosystems, hydrological connectivity, hydrological monitoring, Peace-Athabasca Delta, river flooding, water isotope tracers

1 | INTRODUCTION

Hydrological monitoring in complex, dynamic, water-rich floodplain landscapes is logistically challenging, especially in vast, remote areas at northern latitudes. Overcoming these barriers is essential as these inland freshwaters are globally important yet threatened by climate change and upstream human developments (Dudgeon et al., 2006; Gleick, 2003; Woodward, Perkins, & Brown, 2010). In western North
Recently, Remmer, Kletm, Wolfe, and Hall (2018) used an array of datasets, including measurements of water isotope composition and water chemistry collected one year after a major ice-jam flood event in 2014, as well as paleohydrological records, to assess the effects of flooding on lake conditions in the PAD. They found that the 2014 flood event had short-lived, mostly within-year effects on water balance of the flooded lakes, which they attributed to a diminishing influence of river floodwaters as a consequence of cumulative and unrelenting effects of climate change. Given the importance of spring ice-jam floods and their declining influence during recent decades, tracking the occurrence and characterizing the effects of flood events that do occur is crucial to monitoring efforts and design of potential mitigation strategies. The extent of prior flood events in the PAD has typically been estimated from aerial surveys (Straka & Gray, 2014) and, on occasion, satellite imagery (e.g., Pavelesky & Smith, 2008; Toyrä & Pietroniro, 2005). But, because of the dynamic and somewhat unpredictable nature of ice-jam flood events, delineation of flood extent from aerial observations is highly dependent on the timing and location of flight paths and absence of obstructions that impair detection of river floodwaters into lakes and vegetated terrain (e.g., cloud cover). Neither of these approaches can readily be used to distinguish magnitude of flooding across a landscape nor distinguish different sources of water that may have entered lakes such as river floodwaters versus snowmelt or rainfall. There remains a clear need for a systematic, measurable, and in situ approach to track spatial extent, as well as magnitude, of river flood events on lakes of the PAD and their effect on lake water balances.

Landscape-scale monitoring and assessment of lake hydrological conditions using water isotope tracers has proven to be an effective approach in many remote, northern, water-rich locations (e.g., Brock, Wolfe, & Edwards, 2002; Gibson & Edwards, 2002b; MacDonald et al., 2017). In northern floodplains, this approach is particularly well-suited to partition the relative roles of important hydrological processes on lake water balances because of the strong sensitivity of lake water isotope compositions to influence of isotopically depleted river water versus isotopic enrichment caused by evaporation. For example, water isotope data have been used to estimate the spatial extent of flood events in the PAD (Wolfe et al., 2008) as well as the Slave River Delta located farther downstream within the Mackenzie River Basin (Brock, Wolfe, & Edwards, 2008). In these studies, spring river floodwater dilution to lakes deemed to have flooded was estimated from the difference between the spring and previous fall measurements of water isotope composition. The modelling approach assumed that the spring isotopic depletion was entirely due to the influence of river floodwaters even though there were also likely contributions from snowmelt and/or rainfall. Nonetheless, the flood extent maps utilizing water isotope tracers aligned well with other evidence and demonstrated the usefulness of collection of lake water samples across these landscapes to obtain this information.

Here, our objective is to delineate the extent and magnitude of a spring ice-jam flood event in the PAD that occurred in late April and early May 2018 using water isotope tracers, supplemented by measurements of specific conductivity and field observations. Improving upon previous approaches utilized in the PAD (Wolfe et al., 2008) and Slave...
River Delta (Brock et al., 2008), which did not account for influence of precipitation, we set the water isotope data against an isotope framework established from 16 years of meteorological conditions and online resources, and then develop a set of landscape-specific binary mixing models that allow estimation of the proportion of input to flooded lakes attributable to river floodwater and precipitation (snow or rain). These results demonstrate the value of systematic water sampling and isotope analysis for monitoring and assessment of lake hydrological conditions in dynamic floodplain landscapes such as the PAD.

2 | METHODS

2.1 | Study area

The PAD is episodically fed in the north by the Peace River and more frequently in the south by the Athabasca River (Figure 1). This results in distinct sectors with lakes possessing wide-ranging water balances largely depending on the relative influence of river water. The northern Peace sector (to the north of PAD 37 [Jemis Lake]) is a relic delta that receives floodwater only during infrequent high-elevation spring ice-jam flood events. Consequently, lakes are typically isolated or perched (i.e., closed-drainage) and strongly influenced by evaporation (PADPG, 1973; Pietroniro et al., 1999; Wolfe et al., 2007). Lakes in the southern Athabasca sector (PAD 37 and areas to the south) span a broad gradient of hydrological conditions from those that frequently receive river water in the active delta regions (i.e., restricted-drainage) during both the spring ice-jam and open-water seasons to those that are more substantially influenced by evaporation (i.e., closed-drainage). Broad shallow lakes in the central, low-lying portion of the delta receive continuous river inflow (i.e., open-drainage). Groundwater is considered to be a negligible component of lake water balances in the PAD due to discontinuous permafrost, low hydraulic conductivity of flood-deposited fine-grained sediment (clay and fine silt) that line basins, and low horizontal gradients between lakes (Nielsen, 1972; Prowse, Peters, & Marsh, 1996; Wolfe et al., 2007).

2.2 | Conditions during and preceding the 2018 flood event

Ice-jam flooding occurred along stretches of the Athabasca, Peace, and Slave rivers in late April through early May 2018. For the Peace River, extensive ice-jams developed ≥ ~500 km upstream of the PAD (e.g., Town of Peace River, Sunny Valley, Vermilion Rapids) during April 26–29, 2018, as well as in the Slave River during May 2–4, 2018, at a location ~36–82 km downstream of the northeastern edge of the PAD (Jasek, 2019). Prior to and during this period, ice cover on the Peace River thermally degraded along the reach adjacent to the PAD and was too weak to generate dynamic breakup and substantial overland flooding in the northern Peace sector. However, backwaters from the downstream ice-jam on the Slave River, combined with arrival of a “jave” (ice-jam-release waves; Beltaos, 2008) from release of an upstream ice-jam at Vermilion Rapids, raised river levels at the northern edge of the PAD by 3–4 m and caused flow reversals into the main PAD channels (Rivière des Rochers, Revillon Coupé, Chenal
des Quatres Fouches) during May 3–6, 2018 (Jasek, 2019). Extensive ice-jams formed in the Athabasca River and distributaries (Embarras River, Mamawi Creek; Figure 1) within the southern Athabasca sector between at least May 1–3, 2018, when observations were made during aerial surveys by Kevin Timoney (Tree Line Ecological Research) and Queenie Gray (WBNP). The flooding had begun before the first flight on May 1st and had not ended by the last flight on May 3rd (K. Timoney, Q. Gray, personal communication, May 14, 2019). Maps of the spatial extent of the floodwaters have not yet been produced from the aerial surveys. However, observations and photos suggest overland flooding was extensive in the southern Athabasca sector of the PAD, but limited to a few low-lying basins adjacent to the main channels in the northern Peace sector (K. Timoney, personal communication, May 14, 2019).

Snow cover had melted in the PAD about 2 weeks before the spring floods of 2018 and almost 1 month before our lake and river sampling (May 15–17), as a consequence of consistent above freezing temperatures that occurred after April 19, 2018 (Alberta Climate Information Service https://agriculture.alberta.ca/acis/alberta-weather-data-viewer.jsp, accessed May 15, 2019). Aerial photos provided by K. Timoney identify that most of the sampled lakes became ice free just before the spring flooding, because they show shallow lakes were ice free but deeper lakes (>2 m) still had some remnant ice cover on May 1, 2018, at the time of flooding. A total of 4.1 mm of precipitation (rain) was recorded at Fort Chipewyan (58.7196° N, 111.1407° W) between April 19 and May 17, 2018, a period that precedes river flooding and extends until the end of our lake and river sampling (Alberta Climate Information Service https://agriculture.alberta.ca/acis/alberta-weather-data-viewer.jsp, accessed May 15, 2019).

### 2.3 | Data collection

During May 15–17, 2018, a set of 68 lakes and 9 river sites spanning the range of hydrological conditions across the PAD were sampled with the aid of a helicopter (Figure 1). Water samples and in-situ measurements were collected from a depth of ~10 cm in a mid-lake location (or mid-channel for the river sites). Samples for oxygen and hydrogen isotope analysis were stored in sealed 30 ml high-density polyethylene bottles, and in-situ measurements of specific conductivity were obtained using a YSI ProDSS sonde. Due to the shallow depth of lakes in the PAD, the lake volumes are generally well-mixed at the time of sampling. Thus, the sampling approach provides measurements representative of the total lake storage. Water isotope compositions of the lake and river samples were measured by Off-Axis Integrated Cavity Output Spectroscopy (O-AICOS) at the University of Waterloo in the aid of a helicopter (Figure 1). Water samples and in-situ measurements of specific conductivity were considered simultaneously values of specific conductivity and field observations. Lakes were designated as flooded if (a) water isotope values were close to or overlapping with the isotopically depleted river water isotope compositions, (b) specific conductivity values were close to the range of the river water values (Peace sector: 164.5 to 176.5 μS/cm; Athabasca sector: 179.5 to 188.5 μS/cm), and (c) there was visible evidence of flooding, such as water colour and turbidity (assessed in situ) similar to the closest river or channel, and flooded lake margins and debris (i.e., logs) that appeared to have been carried in by the recent floodwaters. In the few instances where flood status was less obvious, hydrological conditions of nearby lakes were considered in combination with field observations. These cases are detailed in Section 3.

A previously developed isotope framework representing average conditions during 2000–2015 (Remmer et al., 2018) and a coupled-isotope tracer approach (Yi, Brock, Falcone, Wolfe, & Edwards, 2008) were used to calculate the isotope composition of input water (δ) for each flooded lake (Figure 2; Appendix). As prescribed by the coupled-isotope tracer approach (Yi et al., 2008), all lake water isotope compositions experiencing the same atmospheric conditions will fall on lake-specific evaporation lines terminating at δO (i.e., the limiting non-steady-state isotope composition of a lake approaching desiccation), and their intersection with the Local Meteoric Water Line (LMWL) provides an estimate of δI for each lake. In this way, we were able to estimate the isotope composition of the input water entering lakes from our measurement of the isotope composition of lake water. For flooded lakes, we assume that δI consists primarily of floodwater supplied during the flood event and, to a lesser degree, precipitation received during the same period. We recognize that the δI values may also reflect some signal of input water during prior years, although this is likely to be minimal because of the tendency for large volumes of river water to enter the very shallow lakes of the PAD (typically <1 m maximum depth) when they flood.

A set of four Meteoric Water Line Segments (MWLS) were developed capturing the range of lake input water isotope compositions that result from mixing of either rain or snow with floodwater from the Peace or Athabasca River. Rainfall and snowmelt input includes both precipitation falling directly on the lake surface and runoff from the lake catchment that entered the lake during the period captured by our water isotope samples. Estimates of average ice-free season (May-September) rain isotope composition (δ18O = −14.22‰, δ2H = −112.2‰) and average ice-cover season (November-March; note April and October were not used because they typically include both rain and snow) snow isotope composition (δ18O = −26.02‰, δ2H = −199.0‰) were obtained from the Online Isotopes in Precipitation
**FIGURE 2** Schematic $\delta^{18}O$-$\delta^2H$ diagram illustrating water isotope compositions of two hypothetical lakes, one that has received input water consisting mainly of river floodwater and rain ($\delta_{i,-1}$) and one that has received input water consisting mainly of river floodwater and snowmelt ($\delta_{i,-2}$). Each lake plots along a lake-specific evaporation line. The intersection of lake-specific evaporation lines with Meteoric Water Line Segments (snow-to-river [blue] or river-to-rain [red]) provides an estimate of input water isotope composition ($\delta_i$).

Important features of the LEL include the steady-state isotope composition for a terminal basin ($\delta_{SSL}$), which represents the special case of a lake at hydrological and isotopic steady state in which evaporation exactly equals inflow, and the limiting non-steady-state isotope composition ($\delta^*$), which indicates the maximum potential isotopic enrichment of a lake as it approaches complete desiccation. The LEL is anchored at the mean annual isotope composition of precipitation ($\delta_p$). Parameters used in the isotope mass-balance model to derive $\delta_i$ include the lake water isotope composition ($\delta_L$) and the isotope composition of evaporated vapor from the lake ($\delta_E$). See Appendix for further details.

Values generated by Equations 1–4 represent estimates of the proportion of input water attributable to river floodwaters and snowmelt, and do not necessarily represent the total amount of river floodwater in the lake at the time of sampling because nearly all lakes contained pre-existing water. The one exception is PAD 20, which had desiccated prior to the flood event.

Because the rain and snow end-member isotope compositions are not based on local measurements, but rather derived from a global model, and the values can certainly vary, we consider these end-member values to be the largest uncertainty in our modelling approach. Thus, a sensitivity analysis that incorporated uncertainty in the precipitation end-member isotope compositions was performed to evaluate their influence on flood magnitude estimates. We adjusted rain and snow end-member $\delta^{18}O$ values by $\pm1.08\%o$ and $\pm1.88\%o$, respectively, based on 1 standard deviation (SD) of the monthly values. The "OIPC minus" scenario represents application of SD values subtracted from the rain and snow end-member isotope compositions, whereas the "OIPC plus" scenario represents application of SD values added to the rain and snow end-member isotope compositions. The sensitivity analysis was performed using $\delta^{18}O$ values, because a binary mixing model would produce the same result using $\delta^2H$ values. To visualize flood magnitude and explore spatial patterns, the proportion of input water attributed to river floodwater was interpolated between lakes across the surface of the delta. Spatial autocorrelation...
was assessed using Moran’s I (Anselin, 1995). Interpolation was generated using a multiquadratic radial basis function, a deterministic interpolation technique suitable for environmental monitoring (Rusu & Rusu, 2006), with the function and optimal parameters determined through cross-validation. Analyses were performed using ArcMap 10.6 software.

3 | RESULTS

3.1 | Flood status of lakes

Several lake-water isotope compositions cluster around the river isotope compositions when plotted in $\delta^{18}O$-$\delta^2H$ space, which we interpret to reflect substantial influence of river floodwaters (Figure 3a). In contrast, many lake water isotope compositions also plot along the LEL between $\delta_P$ (isotope composition of weighted mean annual precipitation) and $\delta_{SSL}$ (isotope composition of a terminal lake at hydrological and isotopic steady state) indicating evaporative enrichment, which may have occurred during the spring and previous ice-free seasons in the absence of flooding (Figure 3b). The isotope compositions of non-flooded Peace and Athabasca sector lakes overlap, suggesting the influence of evaporative isotopic enrichment in the absence of flooding was comparable across the delta (Figure 3b). There are a few exceptions (8 of 68 lakes) to our strictly isotope-based interpretation of flood status (described further below). These include some lakes that are positioned along the LEL (PAD 27, 14, 58, and 50; Figure 3a), but other data and observations indicate that they received some river floodwaters in 2018. Other exceptions include lakes that are positioned close to the MWLS (PAD 9, M2, M10, and M5; Figure 3b), but measurements of conductivity and field observations led to interpretation that they did not receive river floodwaters in 2018.

Four of the lakes that were offset from the MWLS were considered flooded based on measurements of specific conductivity and visual observations (Figure 3a, Table 1). Three lakes are in the Peace sector (PAD 14, 50, 58) and one lake is in the Athabasca sector (PAD 27). PAD 50 was highly turbid and field observations noted the water colour matched the nearby Claire River, which was observed to be receiving water from the Peace River. We suspect this lake, which had

**Figure 3** $\delta^{18}O$-$\delta^2H$ graphs showing the water isotope compositions of (a) flooded lakes and (b) not flooded lakes sampled in May 2018. Isotope compositions of rain, snow and the Peace and Athabasca rivers are also shown.
| Lake ID | Water isotope composition | Field notes | Conductivity (μS/cm) | Flooded in spring 2018? |
|---------|----------------------------|-------------|----------------------|------------------------|
| Peace Sector |                          |             |                      |                        |
| R1      | -21.24 -162.94            | Very turbid, light brown | 176.5 | Revillon Coupé        |
| R2      | -21.11 -162.64            | Light brown, very turbid - thick with mud | 173.4 | Peace River           |
| R3      | -21.20 -163.88            | Light brown, very turbid, thick with sediment | 175.4 | Chenal des Quatre Fournes |
| R9      | -20.88 -163.35            | Light brown, very turbid, no visible macrophytes | 304.4 | Claire River          |
| R11     | -20.12 -157.90            | Light brown, very turbid | 164.9 | Rivière des Rochers   |
| PAD 1   | -11.60 -115.17            | Brown water, turbid | 329.5 | No                    |
| PAD 2   | -12.66 -121.08            | Brown water, turbid, water lilies emerging | 264.6 | No                    |
| PAD 3   | -13.88 -129.19            | Dark brown water, turbid, water in lake margins | 304.6 | No                    |
| PAD 4   | -11.73 -116.03            | Greenish brown water, some turbidity | 440.6 | No                    |
| PAD 5   | -10.92 -113.10            | Greenish brown water, clear (low turbidity) | 274.6 | No                    |
| PAD 6   | -11.40 -112.96            | Greenish brown water, turbid | 280.9 | No                    |
| PAD 8   | -16.92 -139.06            | Water colour like river channel (red-brown), high water level | 150.5 | Yes                   |
| PAD 9   | -20.61 -162.32            | Clear (low turbidity), very shallow | 56.0 | No                    |
| PAD 12  | -13.75 -128.58            | Dark brown water, clear (low turbidity) | 297.9 | No                    |
| PAD 13  | -12.96 -122.41            | Dark brown water, clear (low turbidity) | 99.5 | No                    |
| PAD 14  | -15.37 -134.08            | Light brown, highly turbid water | 218.1 | Yes                   |
| PAD 15  | -19.35 -155.46            | Light brown, highly turbid water | 241.2 | Yes                   |
| PAD 16  | -12.62 -120.65            | Greenish brown, turbid | 446.4 | No                    |
| PAD 17  | -12.16 -119.88            | Brown water, very clear (low turbidity) | 184.5 | No                    |
| PAD 18  | -9.39 -104.59             | Greenish blue water, very clear (low turbidity) | 167.6 | No                    |
| PAD 50  | -12.06 -118.23            | Light brown water, high turbidity, silty | 324.8 | Yes                   |
| PAD 52  | -14.65 -134.26            | Greenish brown water, low turbidity | 420.3 | No                    |
| PAD 53  | -10.84 -109.09            | Brown water, low turbidity | 242.3 | No                    |
| PAD 54  | -19.41 -155.82            | Light brown water, high turbidity, flooding obvious | 227.2 | Yes                   |
| PAD 57  | -10.99 -114.59            | Greenish brown water, fairly turbid, no signs of flooding | 381.9 | No                    |
| PAD 58  | -13.44 -127.21            | Light brown water, high turbidity | 250.4 | Yes                   |
| PAD 64  | -20.50 -161.40            | Light brown, very turbid, appears flooded, drowned vegetation, water looks to have come from the west end, no visible macrophytes | 254.6 | Yes                   |
| PAD 65  | -11.17 -117.20            | Moderate turbidity, flooding not evident | 316.9 | No                    |

(Continues)
| Lake ID | Water isotope composition | Field notes | Conductivity (μS/cm) | Flooded in spring 2018? |
|---------|--------------------------|-------------|---------------------|------------------------|
| PAD 72  | -10.80 -114.74           | Water level below high-water mark visible on rocks. No signs of flooding noted | N/A | No |
| PAD 73  | -12.68 -124.42           | Water clear, visible to lake bottom | N/A | No |
| PAD 74  | -13.28 -127.51           | Water clear, visible to lake bottom | N/A | No |
| PAD 75  | -13.73 -126.41           | Water is very clear | N/A | No |
| M15     | -11.52 -114.68           | Brown water, low turbidity | 201.8 | No |
| M16     | -12.86 -123.74           | Brown water, low turbidity | 120.1 | No |
| M17     | -18.70 -153.14           | High turbidity, water in lake margins, water levels high | 219.3 | Yes |
| M18     | -17.85 -148.08           | High turbidity, water in lake margins, water levels high | 226.7 | Yes |
| M19     | -11.69 -117.43           | Brown water, low turbidity | 189.7 | No |
| Athabasca Sector |                     |                     |                     |                     |
| R4      | -18.98 -149.85           | Medium brown, very turbid | 179.7 | Athabasca River |
| R5      | -19.12 -151.60           | Medium brown, fast flowing, high turbidity, bank erosion (to the North) | 188.5 | Embarrass River |
| R6      | -19.16 -151.69           | Medium brown, turbid | 181.3 | Fletcher Channel |
| R7      | -19.25 -152.30           | Turbid, brown | 185.1 | Mamawi Creek |
| PAD 19  | -19.17 -153.33           | High turbidity, high water levels | 199.2 | Yes |
| PAD 20  | -19.40 -154.51           | Obvious flooding, high water level, high turbidity | 203.9 | Yes |
| PAD 21  | -19.30 -154.14           | High turbidity, water throughout lake margins, high water level | 211.7 | Yes |
| PAD 22  | -18.70 -149.79           | Fairly turbid, water throughout lake margins | 209.0 | Yes |
| PAD 23  | -18.65 -151.48           | Fairly turbid, water throughout lake margins | 191.0 | Yes |
| PAD 24  | -18.98 -152.12           | Fairly turbid, water throughout lake margins | 189.5 | Yes |
| PAD 25  | -19.04 -152.33           | Fairly turbid | 180.5 | Yes |
| PAD 26  | -19.25 -154.32           | High turbidity, logs in lake on river side | 217.2 | Yes |
| PAD 27  | -16.25 -137.35           | Fairly turbid, water throughout lake margins, high water level | 227.5 | Yes |
| PAD 30  | -19.14 -153.63           | High turbidity, drowned macrophytes, obvious flooding | 217.7 | Yes |
| PAD 31  | -19.27 -154.25           | High turbidity, drowned macrophytes, obvious flooding | 211.4 | Yes |
| PAD 32  | -18.70 -151.52           | High turbidity, water throughout lake margins and surrounding landscape | 244.8 | Yes |
| PAD 33  | -19.40 -155.02           | High turbidity, water throughout lake margins, high water level | 242.4 | Yes |
| PAD 36  | -19.29 -154.65           | Fairly turbid | 225.2 | Yes |
| PAD 37  | -13.62 -129.40           | Clear water, benthic algal growth | 430.7 | No |
been experiencing water-level drawdown during recent open-water seasons (personal observation), received some river water but the isotopically depleted input water was insufficient to offset the prior evaporative enrichment. Similarly, flooding of PAD 58 and PAD 14 was determined by observation of high turbidity and light brown water colour comparable to the Peace River. Turbid water, high water level, and a flooded lake margin was observed at PAD 27.

Four lakes (Athabasca sector: M2, M5, M10; Peace sector: PAD 9) that plot along the MWLS were classified as not flooded based on measurements of specific conductivity and visual observations (Figure 3b, Table 1). Lakes M2 and M5 had very low turbidity and specific conductivity was lower than expected if river water had entered the basins. We suspect these lakes may be influenced by isotopically depleted snowmelt draining through elevated sand dunes located in their catchments. Lake water at PAD 9 was very shallow and clear, which we attribute to a strong influence of isotopically depleted snowmelt. At M10, turbidity of the water was observed to be low, water levels appeared normal, and lake water colour was comparable to nearby lakes (M8 and M9) that we confidently ascertained did not receive river floodwater.

In total, 44% (30/68) of sampled lakes received river floodwaters based on consideration of the water isotope compositions, specific conductivity, and field observations (Table 1). Nine of 32 (28%) sampled lakes in the Peace sector were classified as flooded, and 21 of 36 (58%) lakes in the Athabasca sector were classified as flooded.

### 3.2 Calculation of proportion of input to lakes attributable to river floodwater

Based on calculation of δ values, the input water to each flooded lake was determined to consist mainly of a mixture of either Peace (north...
of PAD 37) or Athabasca (PAD 37 and to the south) river floodwater and rainfall or snowmelt (Figure 4 and Table 2). Lakes were then assigned to one of four MWLS. Input water for 5 of the 9 flooded lakes in the Peace sector consisted mainly of a mixture of river floodwater and rainfall (Figure 4a). Input water for the other 4 flooded Peace sector lakes consisted of mainly a mixture of river floodwater and snowmelt (Figure 4b). In the Athabasca sector, 2 of the 21 flooded lakes received input water mainly comprised of river floodwater and rainfall (Figure 4c), and the other 19 flooded lakes received input water mainly comprised of river floodwater and snowmelt (Figure 4d).

Flooded lakes in the Peace sector contained a mixture of river floodwater and rain or snowmelt, and flooded lakes in the Athabasca sector were dominated primarily by a mixture of river floodwater and snowmelt. Based on Equations 1 and 2, respectively, the proportion of input water attributable to river floodwater in Peace sector lakes dominated by precipitation in the form of rain ranged from 69% to 98% with an average of 81.8%, and the proportion of input water attributable to river water in lakes dominated by precipitation in the form of snow ranged from 90 to 97% with an average of 93.9% (Table 2). Based on Equations 3 and 4, respectively, the proportion of input water attributable to river floodwater in Athabasca sector lakes dominated by precipitation in the form of rain ranged from 93% to 98% with an average of 95.5%, and the proportion of input water attributable to river floodwater in lakes dominated by precipitation in the form of snow ranged from 82% to 94% with an average of 86.3% (Table 2). The average proportion of input water attributable to river floodwater in the flooded lakes was 87.2% (n = 30), with comparable ranges and averages in the Peace (69–98%, avg = 86.9%, n = 9) and Athabasca sector (82–98%, avg = 87.2%, n = 21; Table 2).

### 3.3 Delineation of flood extent and magnitude

Calculations of the proportion of input to lakes attributable to river floodwater were interpolated across the surface of the PAD using a radial basis function to visualize flood extent and magnitude (Figure 5a). Because all flooded lakes had >60% input water attributed to river flooding, only contours of 60% river floodwater input and greater are displayed to highlight the flooded areas. Patterns of flooding differed notably between the Peace and Athabasca sectors. In the Peace sector, river floodwater input was mainly constrained to lakes located along the Chenal des Quatre Fourches (PAD 54 and 58) and the northern reaches of the Revillon Coupé (PAD 14 and 15). Flooding along the Rivière des Rochers and the Claire River was...
limited to a few sampled lakes (PAD 8 and 50). Lakes M17 and M18, which appear to lie at an interior location, received floodwaters from the Rivière des Rochers via a distributary channel. River floodwater was the dominant input into PAD 64, which is in a flood-prone area adjacent to the north shore of the Peace River. In the Athabasca sector, more extensive flooding occurred. River floodwaters entered lakes along Mamawi Creek and western portions of the Embarras and Athabasca rivers. However, flooding was not evident in lakes at the eastern terminus region of the Athabasca Delta. Note that the mapped flooded regions necessarily incorporate lakes that were not sampled, which is reasonable especially for the low-relief Athabasca sector. A sensitivity analysis incorporating uncertainties in the rain and snow end-member isotope compositions in the binary mixing model equations generated small differences in the flood map magnitude estimates (average absolute difference for the OIPC minus scenario = 2.5% (minimum = −5.1%, maximum = 5.5%), Figure 5b;
average absolute difference for the OIPC plus scenario = 2.4% (minimum = −4.6, maximum = 2.9%), Figure 5c).

4 | DISCUSSION

The lateral interaction of rivers and their floodplains is often an important determinant of patterns of biodiversity and species richness (Gregory, Swanson, McKee, & Cummins, 1991; Johnson & Host, 2010; Junk, Bayley, & Sparks, 1989; Ward, 1998; Welcomme, 1979), and monitoring of hydrological connectivity should be considered a vital component of efforts to conserve the ecological integrity of floodplain landscapes. However, measurement of hydrological connectivity can be difficult to obtain in water-rich landscapes, especially during short-lived, episodic flood events (Lindenmayer et al., 2008). Ice-jam floods provide a period of increased hydrological connectivity among rivers, channels, lakes and wetlands of the PAD, and decline in their frequency has remained a prominent concern for decades in recognition that these flood events are critical to maintaining ecological integrity. Thus, it is imperative to track flood events and their influence on hydrological connectivity when they occur, even if some of their effects are short-lived (see Remmer et al., 2018). Historically, this has been done mainly by observation from aircraft, and methods have recently been developed to map flood extent from satellite imagery (Pavelsky & Smith, 2008; Töyrä & Pietroniro, 2005). However, neither of these approaches readily allow quantification of flood magnitude, and the mapped flood extents can be strongly influenced by the timing of observations.

Recently, an international organization has identified the need to expand the scope of monitoring in the PAD, among other recommendations, which has translated into an Action Plan for WBNP (WBNP, 2019; WHC/IUCN, 2017). Here, we demonstrate the utility of systematic sampling of lakes and rivers for analysis of water isotope composition to serve this knowledge requirement, and in particular to quantitatively delineate extent and magnitude of flood events on lakes across the ~6000 km² expanse of the delta. This opportunity arose because, fortuitously, a substantial ice-jam flood event occurred in the midst of our multi-year program of sampling of lakes in the delta, which is designed to generate a monitoring framework to assess changes in hydroecological conditions and sources, distribution and toxicity of contaminants in lakes of the PAD. Had we not been systematically collecting samples prior to and during this time (and engaged in the associated fieldwork planning required to execute the sampling), it would have been substantially more difficult to capture this event in the level of detail presented here, especially if timing and frequency of sampling was sporadic and relied upon by impromptu decision making.

Ice-jams in late April and early May 2018 resulted in flooding of lakes in both sectors of the PAD, although flood extent and magnitude differed between the Peace and Athabasca sectors. High water levels in the Peace River resulted in flow reversals along some river channels, including the Claire River, Chenal des Quatre Fourches, Revillon Coupé and Rivière des Rochers (Jasek, 2019). Results from our study identify that these high river levels led to flooding of several lakes adjacent to these channels (e.g., 96% river floodwater input at PAD 54; Figure 4, 5, 6, Table 2), which has long been observed to occur during high water events on the Peace River (e.g., PADPG, 1973; Wolfe et al., 2006). Interpolation indicates that input to lakes in these areas was dominated by river water, and that the proportion of input water attributed to river floodwater decreased with increasing distance from the channels, consistent with independent observations made during aerial surveys indicating there was minimal overland
flooding in the Peace sector due to absence of ice-jams in the reach of the Peace River adjacent to the PAD (Jasek, 2019).

Our isotope-based methods identify considerably greater spatial extent of floodwaters in the Athabasca sector than the Peace sector, consistent with independent observations made during aerial surveys (Figure 5, K. Timoney, personal communication, May 14, 2019). And, results of our methods were supported by photographic evidence and field notes compiled by K. Timoney (Treeline Ecological Research) showing that Blanche Lake (PAD 25), Dagmar Lake (PAD 33), Mamawi Creek Pond (PAD 30), Johnny Cabin Pond (PAD 31), Richardson Lake (PAD 38), Grey Wavy Lake (PAD 40), and Otter Lake (PAD 46) were all flooded by May 1, 2018 (K. Timoney, personal communication, May 14, 2019). In the Athabasca sector, river floodwaters were conveyed primarily by the Athabasca River and Embarras River and entered lakes along the western portion of these rivers. Flooding along the Embarras River and downstream along Mamawi Creek towards Mamawi Lake aligns with paleolimnological evidence and field observations that river floodwaters have been diverted northward along this flow path due to the Embarras Breakthrough in 1982, a natural river avulsion that has since redirected substantial and increasing volumes of Athabasca River flow (via the Embarras River) along Mamawi Creek (Kay et al., 2019; Timoney, 2013; Wolfe et al., 2008). Lack of observed flooding at the eastern terminus region of the Athabasca Delta also aligns with paleolimnological records that indicate less frequent lake flooding in this area since the Embarras Breakthrough (Kay et al., 2019). Given the importance of the Athabasca Delta terminus region to land users, and their concerns about decreasing water levels (IEC, 2018; MCFN, 2014), confirmation that floodwater is not reaching this area, even during the large ice-jam flood of 2018, is an important contribution to conservation dialogues.

Some lakes in the central area of the Athabasca sector received enough river water input to drastically change their lake water balance.

**FIGURE 6** Images of selected flooded lakes and surrounding areas in May 2017 (left; no flooding) and May 2018 (middle, right; substantial flooding)
(PAD 19-21; Figure 6). For example, PAD 20, a lake located along the western edge of the flood extent, had desiccated by fall 2017 following a multi-year trend of declining water levels. This provided opportunity to evaluate our flood magnitude estimates because there was no water balance memory at PAD 20, which can potentially confound interpretation of δ values. When sampled in May 2018, this lake contained 2.4 m of water, of which we attribute 85% to river floodwater input and 15% to snowmelt (Figure 5 and Table 2). This suggests that 0.36 m of the 2.4 m of new water was contributed by snowmelt. Based on records at the Fort Chipewyan meteorological station (FT. Chipewyan RC, obtained from https://agriculture.alberta.ca/acis/alberta-weather-data-viewer.jsp for the period October 1, 2018 through April 18, 2019 (the snowfall period), only 78 mm (0.078 m) of precipitation fell during the snowfall period. But, the land surrounding the lake has extensive tree and shrub cover, which entraps wind-redistributed snow from adjacent treeless areas. To further evaluate this estimate, we compiled lake depth changes between September 2017 (our prior sampling trip) and May 2018 in Athabasca sector lakes that did not flood. Assuming we sampled lakes in the same locations on both field visits, as intended, lake depth changes ranged from −0.04 m (i.e., a drawdown of 4 cm) to +0.59 m. Thus, our estimate of 0.36 m of snowmelt water contribution to PAD 20 is possible as it falls within this range.

Our sensitivity analysis has demonstrated that incorporating reasonable uncertainty of both the snow and rain end-member isotope compositions does little to change the outcome at the scale presented in Figure 5, which is most appropriate for ecosystem management purposes. Indeed, the most important delineation in our approach is simply whether the lake flooded or not, which water isotope tracers, supplemented by measurements of specific conductivity and field observations, effectively identify. Nonetheless, surveys of rain and snow isotope compositions could better constrain these end-members and be used to assess for spatial heterogeneity, thereby enhancing the accuracy of the model and estimates of flood magnitude.

We acknowledge other uncertainties in this modelling approach. For example, flooded lakes likely contain a mixture of river floodwater, plus rain and snow, and this may have led to over-estimating river floodwater contributions because the linear positioning of these end-members in δ18O-δ2H space precludes the use of a three-component mixing model. We contend that our approach, which considers the two dominant inputs to each lake, provides a reasonable estimation to track the extent and magnitude of river flood events. Indeed, presentation of Figure 5 to WBNP staff indicated that our results largely aligned with observations made during their flyovers (Q. Gray, personal communication, July 20, 2018), and flooding was observed at several lakes along the northern and eastern peripheral flood margins based on our approach (i.e., PAD 30, 31, 38, 40, and 46; K. Timoney, personal communication, May 14, 2019), as stated above. However, none of the lakes identified as flooded by our approach were estimated to have less than 60% of their input attributed to river floodwaters, even those at the outer margin of the flooded area. There are at least two possible explanations for this outcome. One is that our methods are not sufficiently sensitive to detect flooding when this value is less than 60% due to insufficient change in water isotope values, specific conductivity, and other sources of information. The second is that flooding to these shallow, small volume lakes rarely ever results in less than 60% of the input attributable to river floodwater because once rivers overbank, they supply so much water to the flat terrain that more than half of the lake input is due to this hydrological process. Clearly, this suggests a need for further refinement and testing of our methods and more data acquisition to distinguish these possibilities. Also, interpolation of the calculated proportion of input water attributable to river floodwater is limited in some areas of the delta (e.g., Claire River and the northeastern area of the Athabasca sector) by a low density of sampled lakes. Although more lake sampling sites could be added, the set of 68 sampled lakes provides considerable coverage for tracking the main patterns of flooding and hydrological connectivity across the landscape. Despite these uncertainties, the flood map generated by our approach represents a marked improvement over other approaches because it is less prone to logistical limitations (i.e., sampling can occur after a flood occurs, not at the exact moment of flooding) and can quantify spatial variation in the magnitude of flood influence on lakes.

5 CONCLUSIONS AND RECOMMENDATIONS

River water entered lakes in the PAD during an ice-jam flood event in late April to early May 2018. Here we used water isotope tracers, supplemented by measurements of specific conductivity, and field observations, to delineate the extent and magnitude of flooding. We present an isotope-based approach, incorporating a landscape-specific set of binary mixing models that allow estimation of the proportion of input water to lakes attributable to river floodwater and rainfall or snowmelt. Nearly half of the 68 lakes we have sampled since 2015 as part of developing the foundation of a monitoring program received river floodwaters in spring 2018, and the estimated proportion of lake input provided by river floodwater to the flooded lakes averaged 87.2%. We found that flooding in the northern Peace sector was primarily limited to lakes along the Chenal des Quatre Fourches and the northern reaches of the Revillon Coupé. Flooding in the Athabasca sector was much more extensive, but did not reach the eastern terminus of the Athabasca Delta as floodwater was conveyed northwards through the Embarras River and into Mamawki Creek, a continuing legacy of the Embarras Break through that occurred in 1982 (Kay et al., 2019; Timoney, 2013; Wolfe et al., 2008).

Monitoring data are an essential component of science-based policy, and long-term records are the most likely to contribute to policy (Hughes, Beas, Barner, & Brewitt, 2017). However, obtaining resources to support continuous long-term monitoring remains a challenge (Lovett et al., 2007). Cost-effective measurements of important variables can increase the likelihood of the long-term sustainability of a monitoring program, allowing them to survive during times of lean budgets and, thus, increase the probability that the data will contribute to policy and decision making. This is particularly true in remote northern landscapes, which present additional logistical and financial challenges.
(Mallory et al., 2018). For example, conventional water gauges are expensive to install and maintain at a landscape scale and are frequently damaged or lost during the flood event they aim to capture (e.g., Jasek, 2019). Aerial surveys of flood extent can be informative, but are limited by the timing of the flights. In complex and dynamic floodplain landscapes such as the PAD, hydrological monitoring approaches must be sensitive, timely, flexible and cost-effective. Use of water isotope tracers meets these requirements and, thus, provides an alternative and complementary approach capable of capturing an integrative signal of recent hydrological events and processes. Water isotope tracers have a high level of dexterity, sample collection is straightforward and fast (the 77 sites used in this study were sampled over 2 days) and analytical costs are low. Indeed, senior authors of this paper have long utilized water isotope tracers and other approaches in partnership with other northern National Parks in Canada to assist with meeting their aquatic ecosystem monitoring and reporting obligations. Here we propose water isotope tracers along with measurements of specific conductivity as practical, affordable long-term lake hydrological monitoring tool capable of generating informative results for the PAD, and other remote lake-rich landscapes.

ACKNOWLEDGMENTS
C. R. R. was supported by an NSERC Alexander Graham Bell Canada Graduate Scholarship-Doctoral, a W. Garfield Weston Award for Northern Research-Doctoral, Queen Elizabeth II Graduate Scholarship in Science and Technology, and a University of Waterloo President’s Graduate Scholarship. Research was supported mainly by an NSERC Collaborative Research and Development grant and partners (Alberta Environment and Parks, BC Hydro, Canadian Natural Resources Limited, Suncor Energy, Wood Buffalo National Park), with supplemental funding from the Polar Continental Shelf Program and the Northern Scientific Training Program. We thank WBNP staff for logistical support, and Kevin Timoney, Martin Jasek, and Queenie Gray for providing local observations of ice-jam locations in spring 2018.

CONFLICTS OF INTEREST
We declare no conflicts of interest.

DATA AVAILABILITY STATEMENT
The data that support the findings of this study are available from the corresponding author upon reasonable request.

ORCID
Casey R. Remmer https://orcid.org/0000-0003-2626-3012

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APPENDIX A: Meteorological calculations

Relative humidity ($h$) and temperature ($T$) were calculated as the average (monthly) evaporation-flux-weighted values during the period of 2000–2015 using climate data from Environment Canada and the National Research Council of Canada. The average ice-free season $h$ and $T$ were flux-weighted based on potential evapotranspiration determined by Thornthwaite (1948):

$$h_{\text{flux}} = \sum \left( \frac{h \times E_t}{E_t} \right)$$  \hspace{1cm} (A1)

$$T_{\text{flux}} = \sum \left( \frac{T_a \times E_t}{E_t} \right)$$  \hspace{1cm} (A2)

where $T_a$ is the monthly average temperature and $E_t$ is the monthly total evaporation. $E_t$ was calculated using the equation:

$$E_t = 1.6 \frac{L}{D} 30 \left( \frac{10 \cdot T_a}{I} \right)$$  \hspace{1cm} (A3)

where $L$ is the average day length in hours a month, $N$ is the number of days in the month, $I$ is the annual heat index, and $a$ is a calculated coefficient. $I$ was calculated as the total of each ice-free month using the equation:

$$I = \sum I_i$$  \hspace{1cm} (A4)

and $I_i$ was calculated using the equation:

$$I_i = \left( \frac{T_a}{9} \right)^{1.5}$$  \hspace{1cm} (A5)

and $a$ was calculated using the equation:

$$a = 0.49 + 0.0179 \cdot I - 7.71 \cdot 10^{-5} \cdot I^2 + 6.75 \cdot 10^{-7} \cdot I^3$$  \hspace{1cm} (A6)

APPENDIX B: Isotope framework

Isotope framework parameters were calculated (in decimal notation) using approaches described in Gonfiantini (1986), Gibson and Edwards (2002b), Edwards, Wolfe, Gibson, and Hammarlund (2004), and Yi et al. (2008), which are based on the linear resistance model of Craig and Gordon (1965).

The LEL for the PAD region was determined using a 16-year average of environmental conditions described above, as well as pre-existing isotopic data and calculated evaporation-flux-weighted terms. The LEL was determined as a regression of the isotope composition of precipitation ($\delta_p$), the isotope composition of a terminal basin at steady state ($\delta_{SSL}$) and the theoretical limiting non-steady-state composition of a water body approaching complete desiccation ($\delta^*$). $\delta_p$ was obtained from the Online Isotopes in Precipitation Calculator (Bowen, 2016; Bowen & Revenaugh, 2003; IAEA/WMO, 2015). $\delta_{SSL}$ was determined from the average isotope composition of PAD 18, a terminal basin at isotopic and hydrologic steady state located in an elevated portion of the PAD (Yi et al., 2008). $\delta^*$ was calculated from Gonfiantini (1986):

$$\delta^* = \frac{h \cdot \delta_p + \epsilon_h + \epsilon_v / \alpha^*}{h \cdot \epsilon_h + \epsilon_v / \alpha^*}$$  \hspace{1cm} (A7)

where $h$ is the relative humidity (see E.1), $\delta_p$ is the isotope composition of atmospheric moisture for the ice-free season (Gibson, Edwards, & Prowse, 1999), $\epsilon_h$ is the kinetic separation factor between liquid and vapour phases, $\epsilon_v$ is the equilibrium separation factor between liquid and vapour phases, and $\alpha^*$ is the equilibrium liquid-vapour isotope fractionation factors (Horita & Wesolowski, 1994). $\alpha^*$ for $\delta^*H$ and $\delta^*^{18}O$ was derived from the equations described by Horita and Wesolowski (1994):

$$1000 \ln \alpha^* = 1158.8 \left( \frac{T^3}{10^5} \right) - 1620.1 \left( \frac{T^2}{10^3} \right) + 794.84 \left( \frac{T}{10^3} \right)$$  \hspace{1cm} (A8)

$$+ 2.9992 \left( \frac{T^9}{10^5} \right) - 161.04$$

for $\delta^*H$ and

$$1000 \ln \alpha^* = -7.685 + 6.7123 \left( \frac{T^3}{10^5} \right) - 1.6664 \left( \frac{T^9}{10^5} \right) + 0.35041 \left( \frac{T^9}{10^5} \right)$$  \hspace{1cm} (A9)

for $\delta^*^{18}O$ where $T$ represents the interface temperature in degrees Kelvin. The equilibrium separation factor between liquid and vapour phases is expressed as:

$$\epsilon_v = \alpha^* - 1$$  \hspace{1cm} (A10)

and the kinetic separation factor between liquid and vapour phases is expressed as
for $\delta^2$H and

$$
e_k = 0.0125 (1 - h) \quad \text{(A11)}$$

$$
e_k = 0.0142 (1 - h) \quad \text{(A12)}$$

for $\delta^{18}$O, where $h$ is the relative humidity (Gonfiantini, 1986). Isotope composition of atmospheric moisture for the ice-free season ($\delta_{AS}$) was calculated using the equation from Gibson et al. (1999): $\delta_{AS} = \left(\frac{\delta_{SSL} - \epsilon^*}{\alpha^* - e_k (1 - h + e_k)}\right)/h$ \quad \text{(A13)}

Results of the isotope framework calculations are reported in Table A1.

**APPENDIX C: Calculation of isotope composition of lake input water**

$\delta_{I}$ was estimated as the intersection of a regression through $\delta_{E}$ and $\delta^*$ and the Meteoric Water Line Segment, utilizing the coupled isotope tracer approach (Yi et al., 2008). $\delta_{E}$ was calculated using the Craig and Gordon (1965) equation:

$$
\delta_{E} = \frac{\delta_{SSL} - \epsilon^*}{\alpha^* - \delta_0 (1 - h + e_k)} \quad \text{(A14)}
$$

Temperature and relative humidity values for the calculation of $\delta_{E}$ were the same as those used in the framework calculation (flux-weighted average; see Table A1).

**TABLE A1** Values used for calculations of the isotope framework, with references and equation numbers where appropriate

| Parameter | Value | Source | Equation |
|-----------|-------|--------|----------|
| $h$ | 66.19 | Environment Canada, Natural Research Council of Canada | Equation (A1) |
| $T$ | 14.33 $^\circ$C | | Equation (A2) |
| $\alpha^*$ ($^{18}$O, $^2$H) | 1.010, 1.09 | | Equations A8, A9 |
| $\epsilon^*$ ($^{18}$O, $^2$H) (%) | 10.33, 91.26 | | Equation A10 |
| $e_k$ ($^{18}$O, $^2$H) (%) | 4.80, 4.23 | | Equations A11, A12 |
| $\delta_{AS}$ ($^{18}$O, $^2$H) (%) | $-26.90, -203.64$ | | Equation A13 |
| $\delta_0$ ($^{18}$O, $^2$H) (%) | $-18.4, -142$ | OIPC | |
| $\delta^*$ ($^{18}$O, $^2$H) (%) | $-4.29, -81.76$ | | Equation A7 |
| $\delta_{SSL}$ ($^{18}$O, $^2$H) (%) | $-9.18, -104.26$ | Reference Lake (PAD 18) | |
| LEL Slope | 4.27 | | |
| LEL Intercept | $-63.46$ | | |
| Rain ($^{18}$O, $^2$H) (%) | $-14.22, -112.2$ | OIPC | |
| Snow ($^{18}$O, $^2$H) (%) | $-26.02, -199.0$ | OIPC | |