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Zero-Flow Dynamics for Headwater Streams in a Humid Forested Landscape

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ABSTRACT

Much of our understanding on temporary headwater streams is from arid and sub-humid environments. We know less about zero-flow periods in humid headwater catchments that experience seasonal snow cover. Our study characterised the temporal and spatial patterns of zero-flow periods for forested headwater streams in a snow-dominated landscape. We used 36 years of streamflow data from 13 headwater catchments within the Turkey Lakes Watershed located on the Canadian Shield in Ontario, Canada, near the eastern shores of Lake Superior. These headwater catchments differ substantially in their number of May–November zero-flow days (0–166 days per year) despite being clustered in a small geographical area with similar geology, physiography and vegetation cover. The catchments also experience similar continental climatic conditions with relatively even precipitation inputs throughout the year (mean annual precipitation of 1210 mm/year). Inter-annual variability in the number of zero-flow days was primarily associated with May–November precipitation and evapotranspiration. Despite the large seasonal snowpacks that form in this region, the amount of snow did not appear to influence the extent of zero-flow periods. We found that between-catchment variability in zero-flow occurrences was related to differences in catchment area and catchment properties typically associated with greater groundwater influence. Our study suggests that occurrences of zero-flows in headwater streams can be highly variable even over small geographical regions and that flow permanence may be more sensitive to spring to fall weather conditions than the influence of snow due partly to the shallow soils typically found on the Canadian Shield.

1 | Introduction

Stream reaches that lack surface flow for some portion of the year are estimated to comprise a majority of channel networks across the globe (Nadeau and Rains 2007; Datry, Larned, and Tockner 2014; Messager et al. 2021). This fundamental change in the channel environment of temporary streams, from presence of surface flow to standing water and dry channels, can have profound influence on biodiversity, organism life cycles, nutrient dynamics and provision of downstream water resources (Meyer et al. 2007; Bretz, Murphy, and Hotchkiss 2023; Datry et al. 2023a; Malish et al. 2023; Courcoul et al. 2024). There are concerns that climate change is decreasing the duration of surface flow in streams, which may have deleterious effects on aquatic ecosystems of these temporary streams, as well as downstream reaches (Jaeger, Olden, and Pelland 2014; Tramblay et al. 2021; Zipper et al. 2021, 2022; Datry et al. 2023a).

Historically, the majority of research on temporary streams was from arid and semi-arid environments (Buttle et al. 2012; Jaeger, Olden, and Pelland 2014; Reynolds, Shafroth, and Poff 2015; Shanafield et al. 2021; Sabathier et al. 2023). More recently, studies have examined controls on temporary stream dynamics across a range of climatic regions including humid environments (Eng, Wolock, and Dettinger 2016; Zimmer and McGlynn 2017; Botter et al. 2021; Sauquet et al. 2021b; Jaeger et al. 2023). These studies generally use one of two methodologies: (1) flow permanence mapping where stream networks are surveyed for the presence and absence of surface flow (e.g., Jaeger et al. 2019; Pate, Segura, and Bladon 2020; Botter et al. 2021; Sando et al. 2022; Dralle et al. 2023), and (2) use of streamflow records from hydrometric stations to document occurrences of zero-flow observations (e.g., Huxter and Van Meerveld 2012; Eng, Wolock, and Dettinger 2016; Reynolds, Shafroth, and Poff 2015; Sauquet et al. 2021b; Rutkowska et al. 2023). Mapping approaches typically provide only a snapshot of flow permanence across a stream network, but can capture complex spatial variability in presence or absence of surface flow, especially for headwater regions (Pate, Segura, and Bladon 2020; Botter et al. 2021). In contrast, use of hydrometric data can resolve temporal dynamics of flow duration but only for the stream reach where the gauge is located. Hydrometric stations are often installed on larger streams and rivers with permanent flow (Krabbenhoft et al. 2022); therefore, we have a dearth of observations from headwater systems, despite headwater streams making up the majority of stream network length (Van Meerveld et al. 2020). In addition, zero-flow observations at hydrometric stations may not reflect actual stream drying, but result from instrument error, flow reversals, or frozen streams (Zimmer et al. 2020; Herzog et al. 2022).

Drivers of stream drying can be broadly categorised as climatic, geologic and anthropogenic factors (Buttle et al. 2012; Costigan et al. 2016; Datry et al. 2023b). Many studies have documented increased zero-flow occurrences in association with periods or regions with low precipitation and high air temperature (e.g., Reynolds, Shafroth, and Poff 2015; Sauquet et al. 2021b; Tramblay et al. 2021; Zipper et al. 2021). In landscapes that have seasonal snowcover, years with smaller snowpacks can be associated with an increase in summer stream drying (Godsey and Kirchner 2014; Sando and Blasch 2015). Another climatic driver of zero-flow observations in cold climates is stream freezing during winter (Buttle et al. 2012). Geology can also interact with climate to influence stream permanence by controlling water storage, transmissivity and drainage network structure. For example, streambed materials will influence the transmissivity of water from stream to subsurface, with coarser materials promoting greater flow loss and increased surface drying (Sando and Blasch 2015). Increased surface flow duration may be found in geological settings that promote water storage in the subsurface or topographic depressions, such as wetlands and lakes, since these landscape features can provide a downstream water source during dry periods (Waddington, Roulet, and Hill 1993; Jensen et al. 2018; Hudson et al. 2021; Sando et al. 2022). However, in some cases groundwater contributions may be insufficient to sustain surface flow (Warix et al. 2021). Finally, anthropogenic factors can also contribute to stream drying (Datry et al. 2023b). For example, water diversions from groundwater pumping, dams and reservoir operations and land cover changes can alter the frequency and duration of zero-flow periods (Benejam et al. 2010; Ficklin et al. 2018; Widén et al. 2021).

Temporary headwater streams provide a critical connection between terrestrial and aquatic ecosystems, and moderate water, energy, nutrient and sediment transfer between these two systems (Richardson and Danehy 2007; Janisch, Wondzell, and Ehinger 2012; Lupon et al. 2019). In forested environments, there are concerns that climate and forest change, such as wildfire and harvesting, are altering the hydrology of temporary headwater streams, as well as the aquatic ecosystems they support (Brooks 2009; Kampf et al. 2021). As mentioned above, despite their widespread presence, temporary headwater streams are less monitored compared to permanent streams and rivers (Van Meerveld et al. 2020); therefore, we lack even basic information on location and flow duration patterns of temporary headwater streams (Buttle et al. 2012; Kampf et al. 2021; Dralle et al. 2023). Forest management approaches for controlling potential effects on aquatic ecosystems typically differ between temporary and permanent streams, with permanent streams often given more protection from harvesting activities (Hansen 2001; Svec, Kolka, and Stringer 2005; Ontario Ministry of Natural Resources 2010); therefore, there is a need to better understand the hydrology of temporary streams, especially under a changing climate where permanent streams may become temporary or vice versa (Brooks 2009; Zipper et al. 2021).

Our study used streamflow records collected between 1983 and 2020 at 13 headwater streams located at the Turkey Lakes Watershed Study (Webster et al. 2021) to document the spatio-temporal variability in zero-flow frequencies in a humid environment influenced by snow. The catchments are co-located within a small geographical area and have similar climate, soil, geology and forest cover. We evaluated the hypotheses that (a) rainfall inputs and evapotranspiration losses (Buttle et al. 2012; Sauquet et al. 2021b) and (b) the size of antecedent snowpack (Godsey, Kirchner, and Tague 2014; Sando and Blasch 2015) control the inter-annual variability in number of zero-flow days. In terms of differences in number of zero-flow days between catchments, we expected catchments with larger drainage areas to have fewer zero-flow days than smaller catchments (Svec, Kolka, and Stringer 2005; Reynolds, Shafroth, and Poff 2015; Sando et al. 2022). We also expected catchments characterised by greater groundwater influence and water storage, such as wetlands, to be associated with fewer zero-flow days (Jaeger et al. 2019; Sando et al. 2022).

2 | Methodology

2.1 | Study Site

The Turkey Lakes Watershed (TLW; 47°03′N, 84°25′W) is located approximately 65 km north of Sault Ste. Marie, Ontario, Canada (Figure 1). The forested watershed is 10.4 km^2 in size and has a total relief of 295 m (Webster et al. 2021). The TLW is located in the Great Lakes St. Lawrence forest region within the Boreal Shield Ecozone (Rowe 1972) and is covered by a mixed hardwood-coniferous forest. Upland sites are dominated by sugar maple (*Acer saccharum* Marsh., 90%), yellow birch (*Betula alleghaniensis* Britton., 9%), and various conifers (1%), whereas lowland sites have a higher proportion of conifers (Jeffries, Kelso, and Morrison 1988). The TLW bedrock is Precambrian metamorphic basalt (silicate greenstone) with granitic outcrops (Hazlett, Semkin, and Beall 2001). The bedrock is overlain by silty to sandy soils that vary in depth from less than 1 m in the headwater areas to 1–2 m at lower elevations (Jeffries, Kelso, and

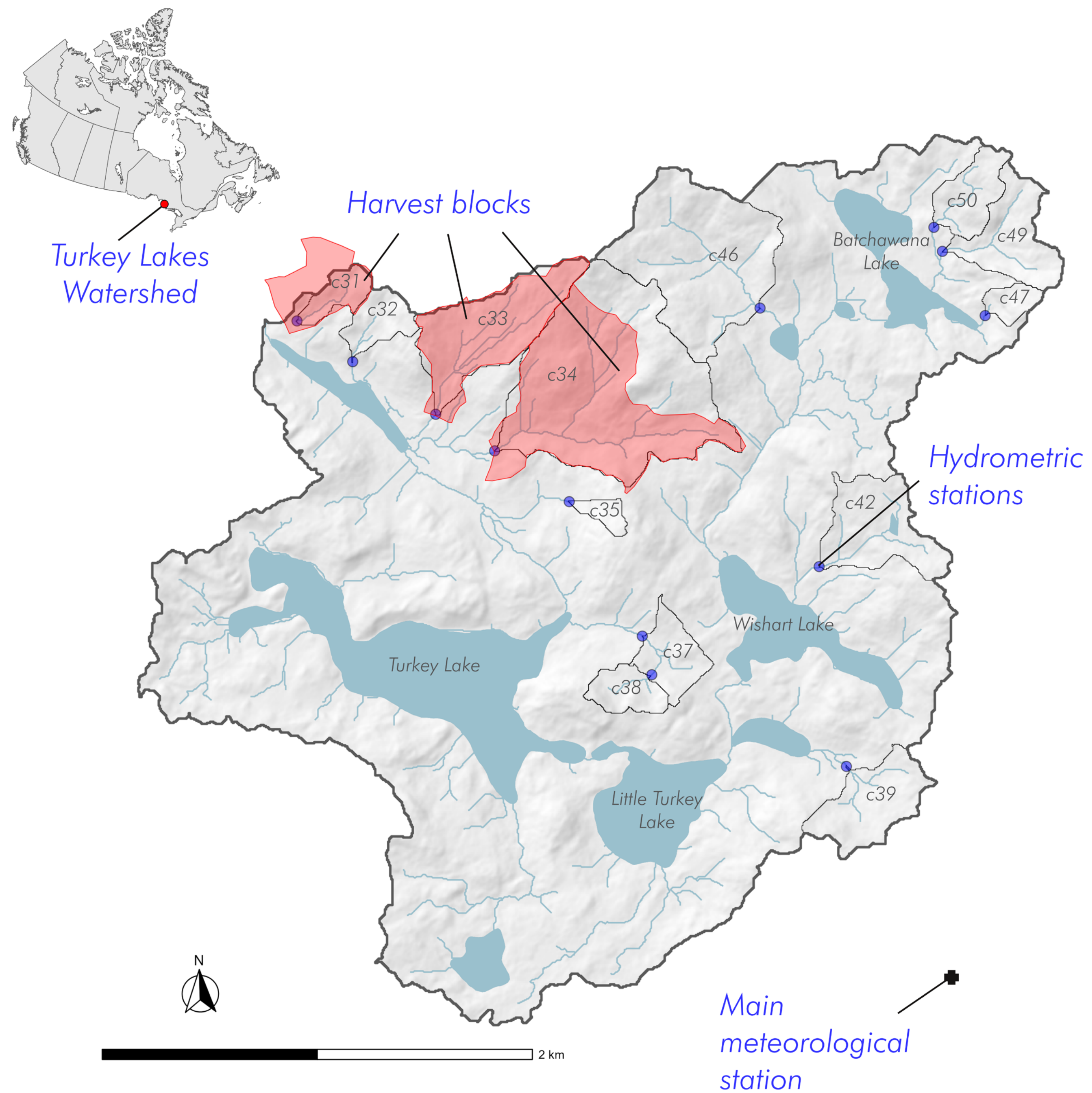

FIGURE 1 | Map of the Turkey Lakes Watershed (TLW). The 13 headwater catchments and their weir locations (blue circles) are shown within the greater Turkey Lakes Watershed. The 1997 forest harvesting is shown in red. Inset map in top left corner situates the Turkey Lakes Watershed within Canada.

Morrison 1988) and organic soil accumulates in bedrock depressions and riparian areas (Creed et al. 2008).

The watershed has a continental climate with significant influence from Lake Superior (Creed et al. 2003) and receives an average of 1210 mm of annual precipitation (based on data collected at the TLW main meteorological station from 1981 to 2021). Approximately 35% of total precipitation is in the form of snow, which begins to accumulate in November and remains until March–May (Beall, Semkin, and Jeffries 2001; Webster et al. 2021). For the period of 1981–2021, the mean annual air temperature was 4.5°C with a January monthly mean of −10.4°C and a July monthly mean of 17.8°C. The TLW forest is undisturbed, with the exception of a light selective harvest in the 1950s (Beall, Semkin, and Jeffries 2001) and an experimental harvest in late summer and fall of 1997 involving three monitored headwater catchments (Figure 1; Table 1; Webster et al. 2022).

2.2 | Zero-Flow Observations

Daily mean streamflow for 1983–2020 from the 13 headwater catchments (Table 1) was estimated from stage-discharge relationships developed using weir equations for v-notch weirs,

TABLE 1 | Properties of the 13 headwater catchments at the Turkey Lakes Watershed.

Catchment	Harvest	Area (ha)	Mean elev (m)	Relief (m)	Wetland cover (%)	Mean slope (%)	Aspect	Mean flow length (m)	MTT (months)	Winter Tw (°C)
c31	Clearcut	4.9	405	80	2.0	25	Southwest	27	20.6	0.9
c32	Reference	6.6	413	107	1.5	30	Southwest	31	18.7	0.7
c33	Selection cut	24.1	470	272	0.4	28	Southwest	45	19.5	0.6
c34	Shelterwood	68.9	474	264	1.2	30	Southwest	44	21.5	0.6
c35	Reference	4.5	447	106	0.0	34	West	67	21.0	1.6
c37	Reference	15.3	401	57	12.4	21	Northeast	31	10.1	0.7
c38	Reference	6.3	416	46	12.4	23	Northeast	22	NA	0.5
c39	Reference	15.6	415	82	1.9	20	North	35	18.0	2.0
c42	Reference	18.3	477	112	6.6	22	Southwest	30	10.0	1.2
c46	Reference	43.0	544	140	1.4	29	Southeast	42	10.3	0.4
c47	Reference	3.5	553	96	0.0	36	West	36	11.9	0.8
c49	Reference	14.9	554	100	2.0	27	Southwest	25	8.2	0.4
c50	Reference	9.2	554	83	7.6	23	Southwest	24	2.8	0.4

Note: Catchment area, mean elevation, relief, mean slope, aspect and mean flowpath length were extracted from a lidar-derived 5 m digital elevation model. Percent wetland cover was based on field measurements outlined in Creed et al. (2003). Harvest refers to whether the catchment was harvested during the 1997 experimental harvesting (Buttle et al. 2018) or was unharvested (reference). Estimated mean travel times (MTT) are from Leach et al. (2020) and details on the mean winter stream temperature (Tw [°C]) measurements can be found in Hudson, Leach, and Houle (2023).

stilling wells and water level loggers (Beall, Semkin, and Jeffries 2001; Buttle et al. 2018). Weirs were installed into the basal till in an effort to capture all flow from the catchments; however, there was the potential for some subsurface flow to bypass the weirs.

The number of days with zero-flow recorded at each catchment outlet between 1 May and 30 November for each year was extracted from the streamflow records. A day was determined to have zero flow if stage measured at the hydrometric station was below the elevation of the v-notch for the entire 24 h period; therefore, days where stage was above the v-notch elevation for only part of the day were not classified as zero-flow days. We focused on the open water period (May–November) since extensive ice formation during the winter months can make flow estimates more uncertain. We recognise that subsurface flow may have by-passed the hydrometric stations, resulting in zero-flow observations despite surface flow occurring upstream and downstream of the gauge (Zimmer et al. 2020). In addition, due to the construction of the weirs, zero-flow observations do not necessarily indicate that the streambed is dry as there may be standing water behind the weirs even though the water level is below the v-notch. We consider these challenges, and potential implications for the key conclusions, in the discussion.

Years with at least 20% missing streamflow data from a specific station were removed from the analysis. In addition, the post-harvest years (1997–2020) for c31, c33 and c34 were not included since the number of zero-flow days may be influenced by harvesting effects (Buttle et al. 2018, 2019).

2.3 | Meteorology

Meteorological measurements were made at a 10 m tower (411 m above sea level) situated about 1.5 km south-east of the TLW catchment boundary (Figure 1). Sensors for air temperature and relative humidity were logged over 10 min intervals and averaged over a 24 h period to provide daily mean values. Sensor types have changed throughout the operation at TLW. Customised air temperature and hygrodynamics H-170 sensors was used from 1980 to 1989 for air temperature and relative humidity, respectfully (Semkin et al. 2012). From 1989 to 2007, a Vaisala HMP35C for both air temperature and relative humidity was used and a Rotronic MP100H has been in operation since 2007. During each change in sensor type, both sensors were run concurrently to cross-check values and ensure consistency in the long-term records. Solar radiation has been measured with an Eppley Precision Spectral pyranometer (Model PSP) and logged at 10 min intervals. Precipitation was summed over a 24 h period to provide daily totals. From 1980 to 2017, accumulation measured by standard rain gauges and Nipher-shielded snow gauges were recorded by field technicians during daily site visits. From 2017 onwards, an OTT Pluvio2 all-phase precipitation gauge has been used. We estimated daily potential evapotranspiration using the approach by Priestley and Taylor (1972). We assumed an albedo of 0.15, which is typical of forest cover dominated by sugar maple trees (Bourque et al. 1995).

Snow water equivalent (SWE) on 1 March was modelled using a daily snow accumulation and melt routine within a hydrologic model calibrated to stream discharge measured at catchment c32. The hydrologic model used was a version of HBV-EC (Hamilton, Hutchinson, and Moore 2000) within the Raven hydrologic modelling framework (Craig et al. 2020). Although the focus of the modelling was to estimate snow conditions, we calibrated the model to mean daily stream discharge since this should provide an integrated signal of both snowmelt timing and interception losses in this snow-dominated hydrologic regime. We calibrated the model to discharge measured at the c32 catchment in part because it is a relatively small catchment whose hydrology should be dominated by vertical water fluxes and it has a long-term snow survey site in close proximity. Details of the model application are provided in Leach et al. (2020). In brief, the model was run at a daily time step from 1980 to 2020 using meteorological data collected at the TLW meteorological station. We calibrated the model to the 1983–2012 period, allowing for a three-year spin-up period (1980–1982). Comparisons of modelled SWE with a nearby snow survey site suggested the model generally captured seasonal snow dynamics (root mean square error of 44 mm). We compared a few different SWE metrics, such as maximum SWE, date of last snow and April 1 SWE and found that March 1 SWE showed a strong correlation with these other metrics (*r* between 0.64 and 0.89).

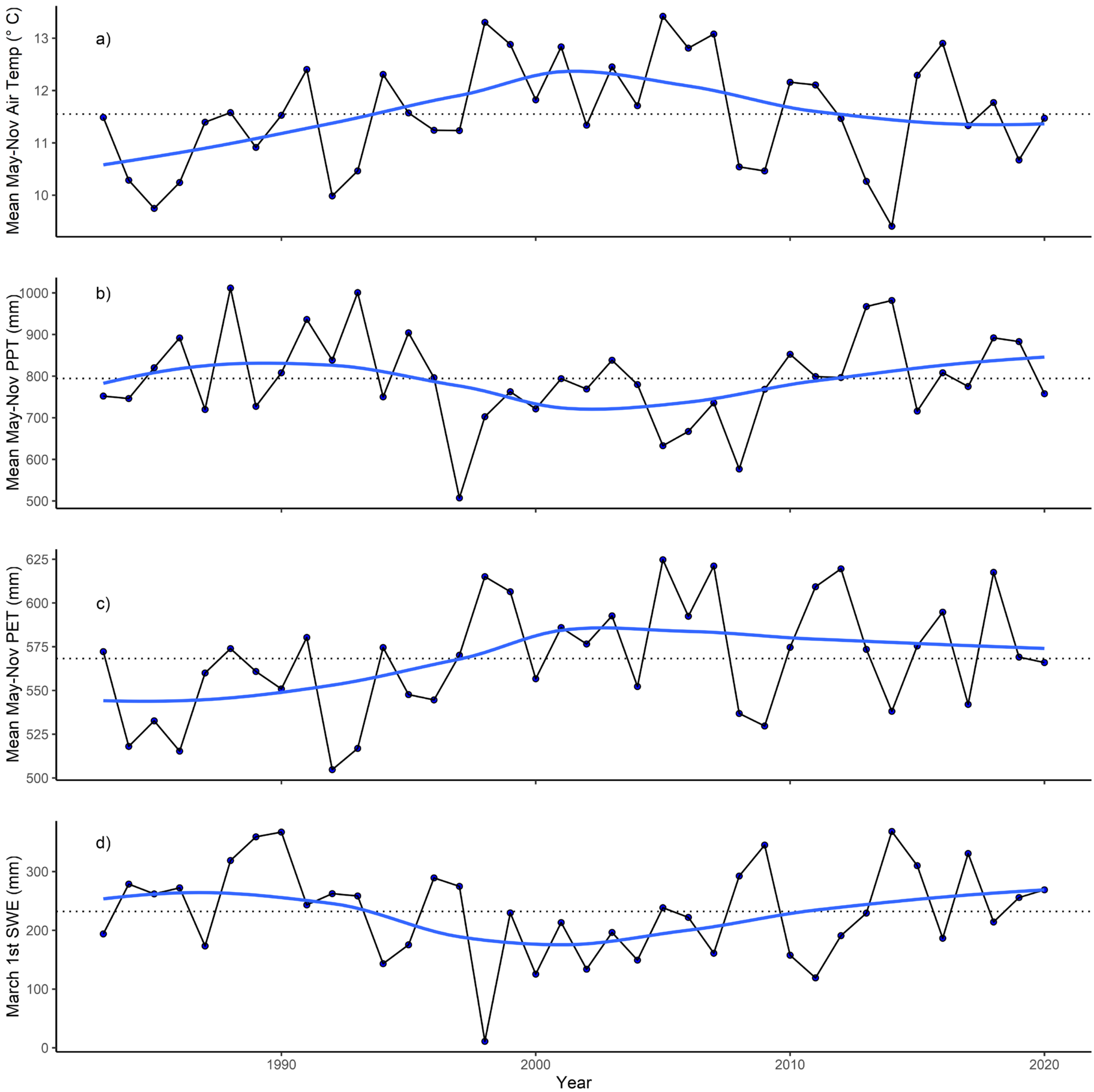

FIGURE 2 | Climatic overview of the 1983–2020 study period. Plot (a) shows mean May–November air temperature (°C). Plot (b) shows total May–November precipitation (mm). Plot (c) shows estimate potential evapotranspiration (PET; mm) for May–November. Plot (d) shows modelled snow water equivalent (SWE; mm) on March 1. Horizontal dotted lines are long-term mean values. LOESS smoothed lines are shown in blue to help visualise long-term trends.

2.4 | Catchment Characteristics

We extracted catchment characteristics from a lidar-derived 5 m digital elevation model of the TLW (Creed et al. 2003) using SAGA GIS (Conrad et al. 2015). The terrain-based characteristics included: catchment area (determined using the D8 method), elevation, mean topographic wetness index (Beven and Kirkby 1979), mean catchment slope and mean hillslope flowpath length. For the hillslope flowpath length, which requires stream channel network locations to know when a hillslope connects to a channel, we assumed a channel initiation threshold of 10 000 m^2 based on knowledge of the channel locations (Webster et al. 2011). Percent wetland cover was based on field measurements reported by Creed et al. (2003). Each catchment was surveyed on foot and the perimeters of surface or near surface-saturated areas were mapped during June 2000. The primary motivation for including elevation is because higher elevation catchments at TLW tend to have shallower soils than catchments at lower elevations (Jeffries, Kelso, and Morrison 1988).

In addition to terrain-based catchment characteristics, we also considered two estimates of groundwater influence on streamflow: (1) mean catchment travel time and (2) winter stream temperature. Mean catchment travel time provides an estimate of the time it takes a water molecule to travel from where it enters soil as rain or snowmelt to when it exits the catchment outlet (McGuire and McDonnell 2006; Soulsby, Tetzlaff, and Hrachowitz 2009). Leach et al. (2020) estimated mean travel times for 12 of the 13 headwater catchments at TLW using chloride as a tracer and a lumped convolution approach using the gamma transfer function (Kirchner, Feng, and Neal 2000). Stream temperature can be an indicator of groundwater influence (Kelleher et al. 2012; Hare et al. 2021). Streams with a greater proportion of groundwater inflow or groundwater sourced from deeper flowpaths would be expected to have elevated winter water temperatures compared with streams with less groundwater contribution or characterised by shallow flowpaths (Anderson 2005). The streams at TLW develop ice cover during the winter; therefore, water temperatures are less influenced by differences in energy exchanges at the stream-air interface compared to summer periods and should allow for better detection of a groundwater influence (Leach et al. 2023). Field technicians measured water temperature bimonthly during winter months using a mercury-filled pocket thermometer (1983–2001; ±0.5°C) or digital thermometer (2001-onwards; ±0.1°C) after augering through the ice. All the streams are sampled on the same day, which helps account for hydrometeorological variations (Leach, Hudson, and Moore 2022; Hudson, Leach, and Houle 2023). From these measurements, we computed long-term mean January and February stream temperature for each catchment.

2.5 | Analysis

We used a combination of graphical and statistical modelling to explore temporal and spatial variability in zero-flow days. To account for the relative influences of snow and meteorology during the open-water period on inter-annual variation in zero-flow days, we considered SWE on 1 March, and total precipitation and potential evapotranspiration for May–November. We also evaluated total precipitation and total potential evapotranspiration from October of the preceding year until November of the year of interest as potential explanatory variables. Our rationale was that antecedent conditions during winter and the preceding autumn might influence the number of zero-flow days; however, these variables provided no additional explanatory power for inter-annual variability in number of zero-flow days. We speculate in the discussion that this may be due to the shallow soils found in the region resulting in antecedent

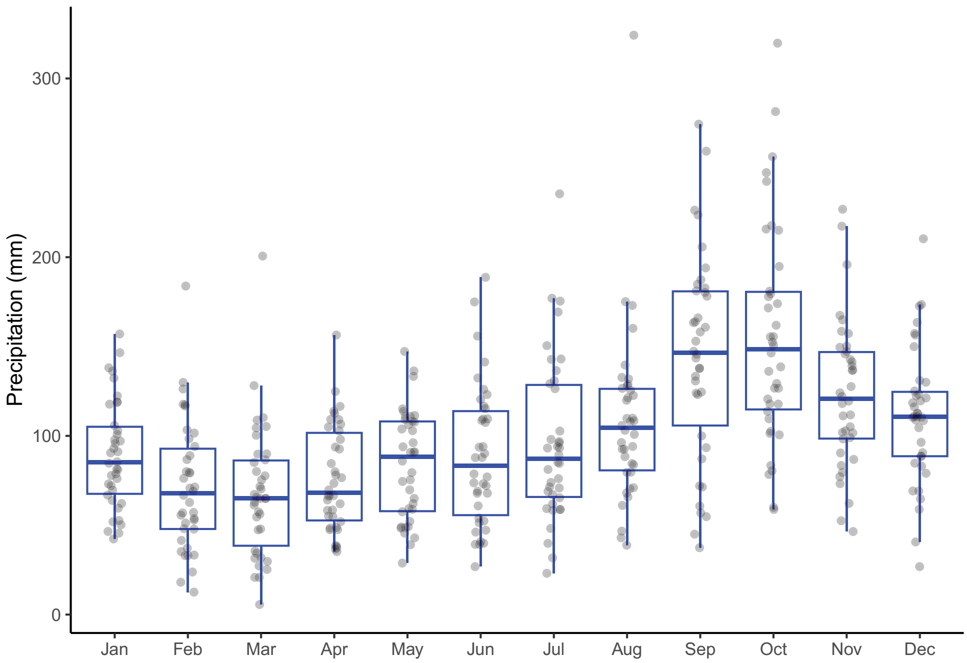

FIGURE 3 | Boxplots of monthly total precipitation (mm) for 1983–2020. The thick horizontal line is the median value, the lower and upper bounds of the box are the first and third quartiles, and the lower and upper whiskers extend from the first and third quartiles to a data point that is no further than 1.5 times the inter-quartile range. Grey points are individual years.

conditions being less influential on the occurrences zero-flow days. Linear regression models were fit to the number of zero-flow days during May–November (NZF) for each catchment:

$$\mathrm{NZF} = \beta_o + \beta_1 \mathrm{PPT} + \beta_2 \mathrm{PET} + \beta_3 \mathrm{SWE} + \varepsilon, \tag{1}$$

where PPT is May–November precipitation (mm), PET is May–November potential evapotranspiration (mm), SWE is the snow water equivalent (mm) on 1 March, β_o, β_1, β_2 and β_3 are estimated coefficients and ε is the error term. The linear models were fit to each catchment and we compared the resulting coefficient estimates and their 50% and 95% uncertainty intervals to infer differences in relationships between the predictor variables and the number of zero-flow days for the catchments. Prior to fitting the models, we standardised the predictor variables by subtracting the mean and dividing by two times the standard deviation so that coefficient estimates were directly comparable (Gelman 2008). We assessed model fits by evaluating residual plots and using R^2 for explained variance.

For the spatial variability analysis, we examined correlations between number of zero-flow days during May–November for a given year and catchment characteristics. We formalised this comparison by fitting a linear mixed-effects model to the number of zero-flow days. Many of the catchment characteristics are correlated; therefore, we only retained four of the eight catchment characteristics in the model based on data availability (e.g., mean transit time estimates were not available for all the streams and were also weakly correlated with number of zero flow days) and reducing redundancy in explanatory variables (e.g., slope, TWI and wetland cover are highly correlated). The final model had the following structure:

$$\begin{aligned}\mathrm{NZF}_j = \beta_0 + b_{0i} + (\beta_1 + b_{1i})\log\mathrm{Area}_j + (\beta_2 + b_{2i})\mathrm{wetland}_j \\ + (\beta_3 + b_{3i})\mathrm{Tw}_j + (\beta_4 + b_{4i})\mathrm{elev}_j + \varepsilon_j,\end{aligned} \tag{2}$$

where NZF_j is the number of zero flow days for stream j, $\log\mathrm{Area}_j$ is the logarithm of catchment area for stream j, $\mathrm{wetland}_j$ is the percent wetland cover for stream j, Tw_j is the mean January–February stream temperature for stream j, and elev_j is the mean catchment elevation for stream j; β_0, β_1, β_2, β_3 and β_4 are fixed-effects coefficients to be estimated; b_{0i}, b_{1i}, b_{2i}, b_{3i} and b_{4i} are random effects for the ratio of May–November precipitation to potential evapotranspiration (PPT/PET) for year i, and ε_j is the error term. We used the ratio of precipitation to potential evapotranspiration (PPT/PET), often referred to as the aridity index (Zipper et al. 2021), to account for inter-annual

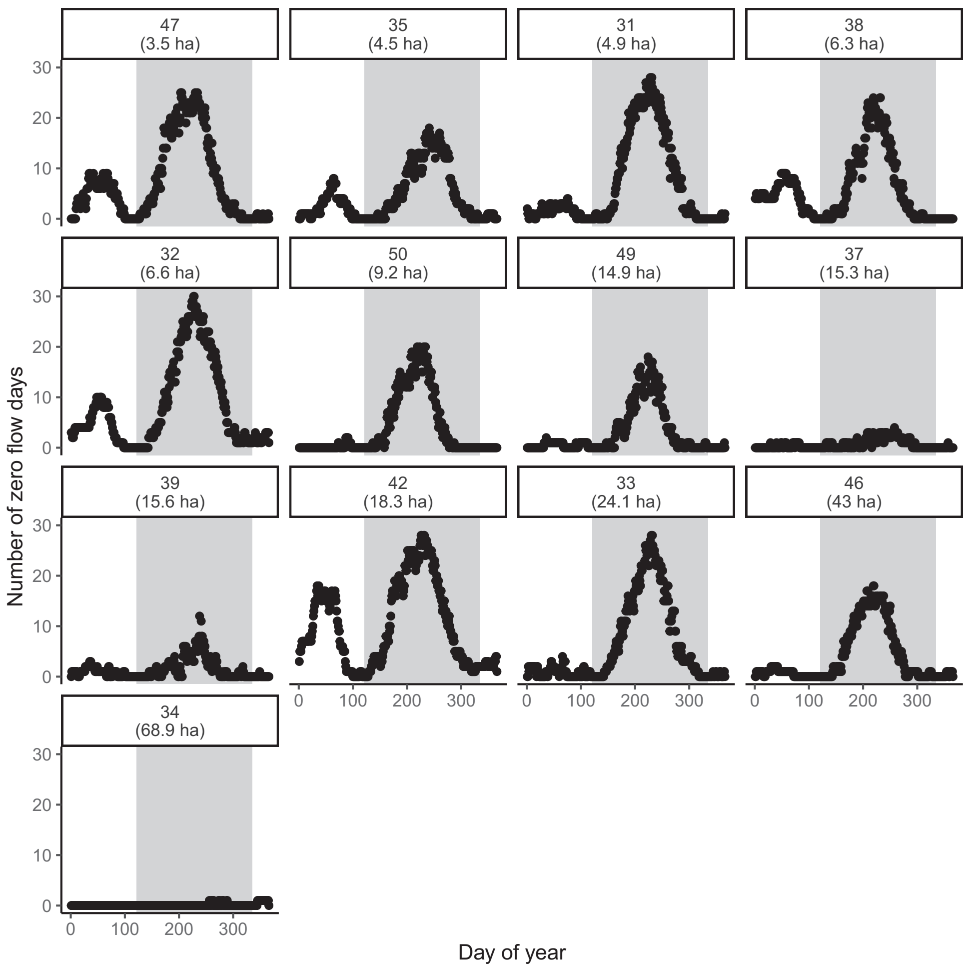

FIGURE 4 | Number of zero-flow days at each study catchment for a given day of the year over the 1983–2020 study period. The grey band shows the May–November period. Individual plots are labelled with the catchment number and the catchment area in parentheses. Plots are ordered by catchment area from top-left to bottom-right. The data for the harvested catchments (c31, c33 and c34) include the post-harvest period (1998–2020) in order to use the same y-axis scale between catchments.

variability in meteorology. Prior to fitting the models we standardised the predictor variables by subtracting the mean and dividing by two times the standard deviation so that coefficient estimates were directly comparable (Gelman 2008). We computed the intraclass correlation to quantify the proportion of variance explained by the random effect (PPT/PET) to the total variance (Hox, Moerbeek, and Van de Schoot 2017). The model was fit using the *lme4* package in R (Bates et al. 2015).

3 | Results

3.1 | Climate Overview

Long-term (1983–2020) mean air temperature and total precipitation during the May–November period was 11.5°C and 794 mm, respectively (Figure 2). Annual variability in May–November mean air temperature ranged from 9.4°C in 2014 to 13.4°C in 2005. Total May–November precipitation was lowest in 1997 (507 mm) and highest in 1988 (1012 mm). Estimated potential evapotranspiration was lowest in 1992 (505 mm) and highest in 2005 (625 mm). Modelled 1 March SWE ranged between 11 mm in 1998 to 368 mm in 2014. In general, the climate was cooler and wetter than average during the first decade of the study period, warmer and drier than average from the mid-990s to about 2010, and returned to long-term average conditions over the last decade. Monthly mean precipitation at the study site ranged between 67 mm/month in February to 155 mm/month in October (Figure 3).

3.2 | Zero-Flow Frequencies

Most of the streams are characterised by prolonged zero-flow periods during the summer, with the exception of c34 and c37 (Figure 4). Streams had comparatively few zero-flow days during the spring freshet and fall periods. Inter-annual variability in number of zero-flow days varies across streams (Figure 5). Stream c34 always had flow during the 1983–1996 period (c34 was harvested in 1997 and data after harvest are not considered here). Streams c39 and c37 had some of the lowest frequencies of zero-flow days and in many years maintained flow throughout the May–November period (48% and 82% of years with at least 80% daily discharge data for c39 and c37, respectively). Streams with the most frequent occurrences of zero-flow days include c32 and c42. These two streams always

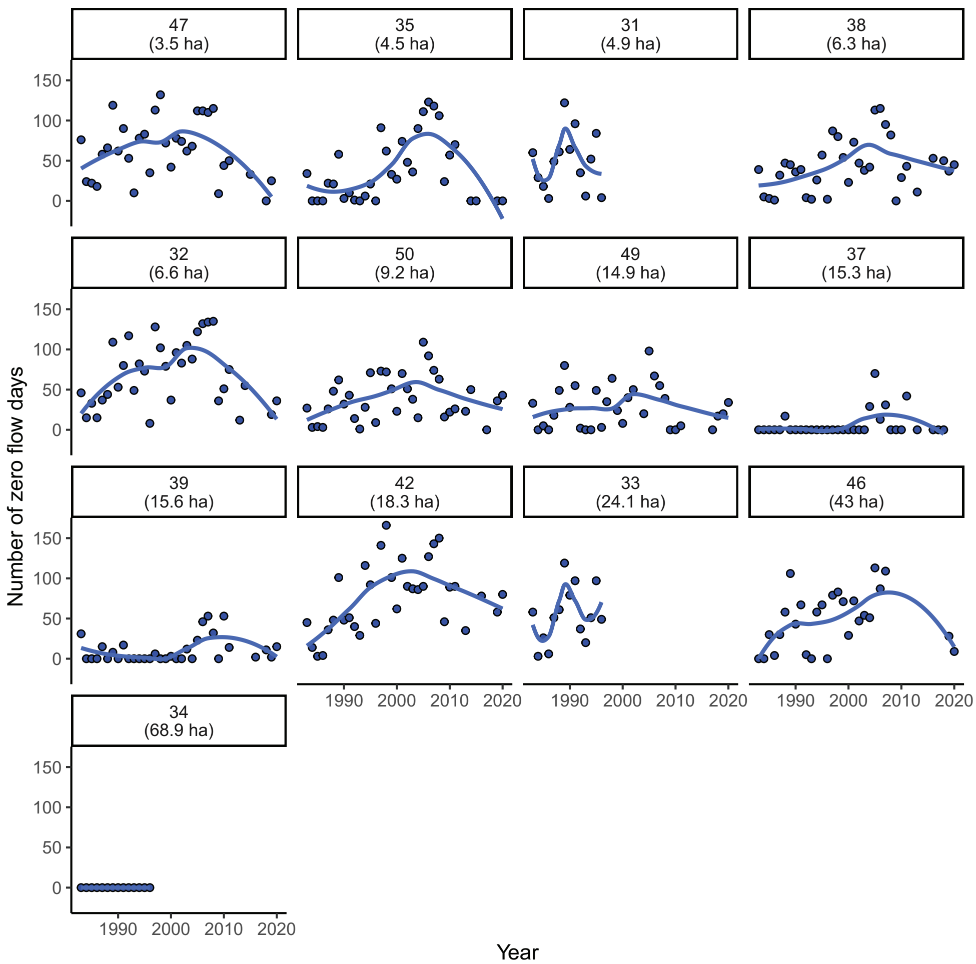

FIGURE 5 | Number of zero-flow days per stream during May–November for each year. LOESS smooth line is shown for visual reference. Only pre-harvest years (1983–1996) are shown for the three catchments that were harvested (c31, c33 and c34). Years with more than 20% missing flow data are not shown. Individual plots are labelled with the catchment number and the catchment area in parentheses. Plots are ordered by catchment area from top-left to bottom-right.

exhibited at least 3 days of zero-flow per year during the study period, with a maximum number of zero-flow days of 135 (c32) and 166 (c42). For the unharvested streams with minimal data gaps, the number of zero-flow days generally increased from 1983 to the mid-2000s, followed by a decreasing trend from the mid-2000s to 2020.

3.3 | Temporal Variability

We wanted to explore whether precipitation, potential evapotranspiration and 1 March SWE could account for the inter-annual variability in number of zero-flow days. Potential evapotranspiration had stronger correlations ($r = 0.38–0.76$; Figure 6) with zero-flow days than precipitation ($r = -0.6$ to -0.05; Figure 7) or 1 March SWE ($r = -0.4$ to 0.12; Figure 8). Coefficient estimates from fitting models (Equation 1) to the 13 streams also show that the number of zero-flow days has a consistent positive (negative) relationship with potential evapotranspiration (precipitation), with the exception of c34, which had no zero-flow days during the period used for analysis (Figure 9). In contrast, 1 March SWE did not show a strong or consistent relationship with number of zero-flow days for any of the streams. The R^2 values for these model fits ranged between 0.18 (c37) and 0.76 (c38), and streams with greater inter-annual variability in number of zero-flow days tended to be associated with higher R^2 values.

3.4 | Spatial Variability

Correlations between number of zero flow days and catchment characteristics during relatively wet years with larger PPT/PET values were typically low (Figure 10). In contrast, some catchment characteristics, such as catchment area, January–February stream temperature, slope, TWI and percent wetland cover, had stronger correlations with zero flow days during relatively drier years. Regardless of PPT/PET status, mean flowpath length, mean transit time and mean elevation were not strongly correlated with number of zero flow days. Coefficient estimates from the mixed-effects linear model highlight that catchment area, mean January–February stream temperature, and percent wetland cover were negatively related to number of zero flow days (Figure 11). The intraclass correlation for the model suggested that interannual variability in PPT/PET accounted for 43% of the total variance.

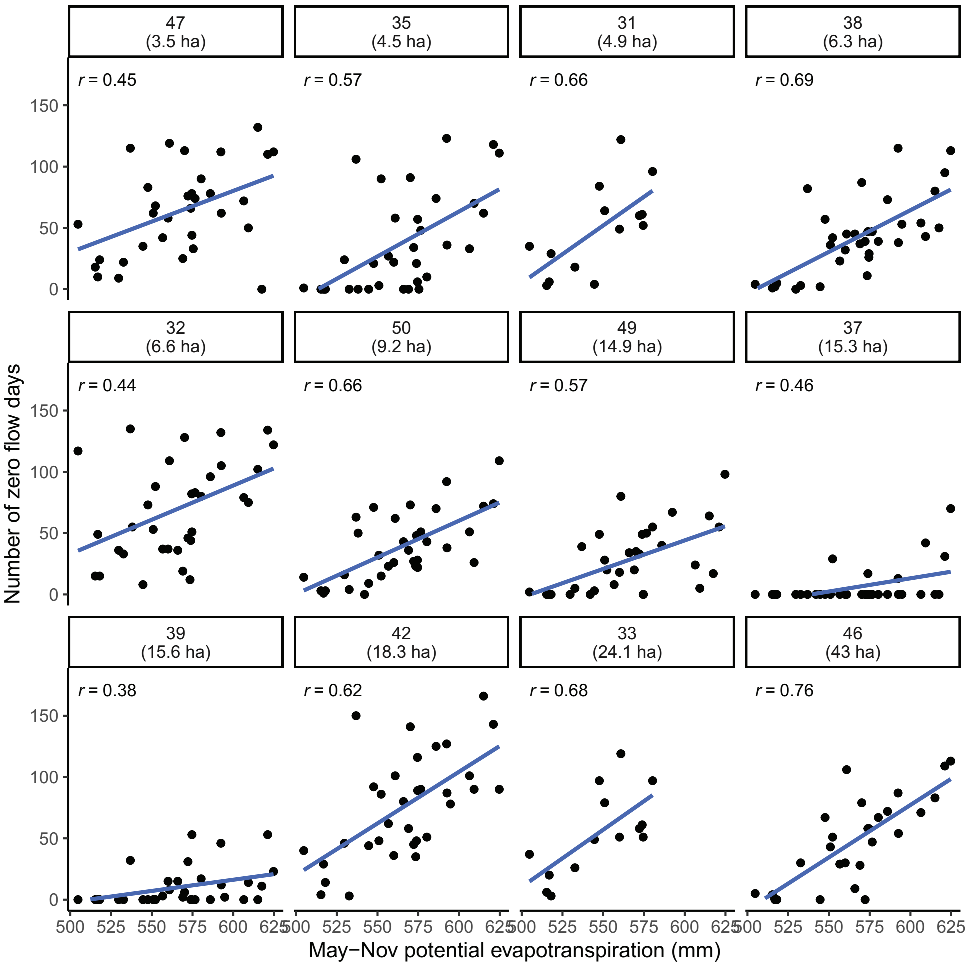

FIGURE 6 | Number of zero-flow days plotted against May–November evapotranspiration for 12 streams. Line of best fit is shown in blue and the Pearson correlations (r) are shown inset. Only pre-harvest years (1983–1996) are shown for the two catchments that were harvested (c31 and c33). Note that c34 is omitted as it did not have any zero-flow days during the pre-harvest period. Years with more than 20% missing flow data are not shown. Individual plots are labelled with the catchment number and the catchment area in parentheses. Plots are ordered by catchment area from top-left to bottom-right.

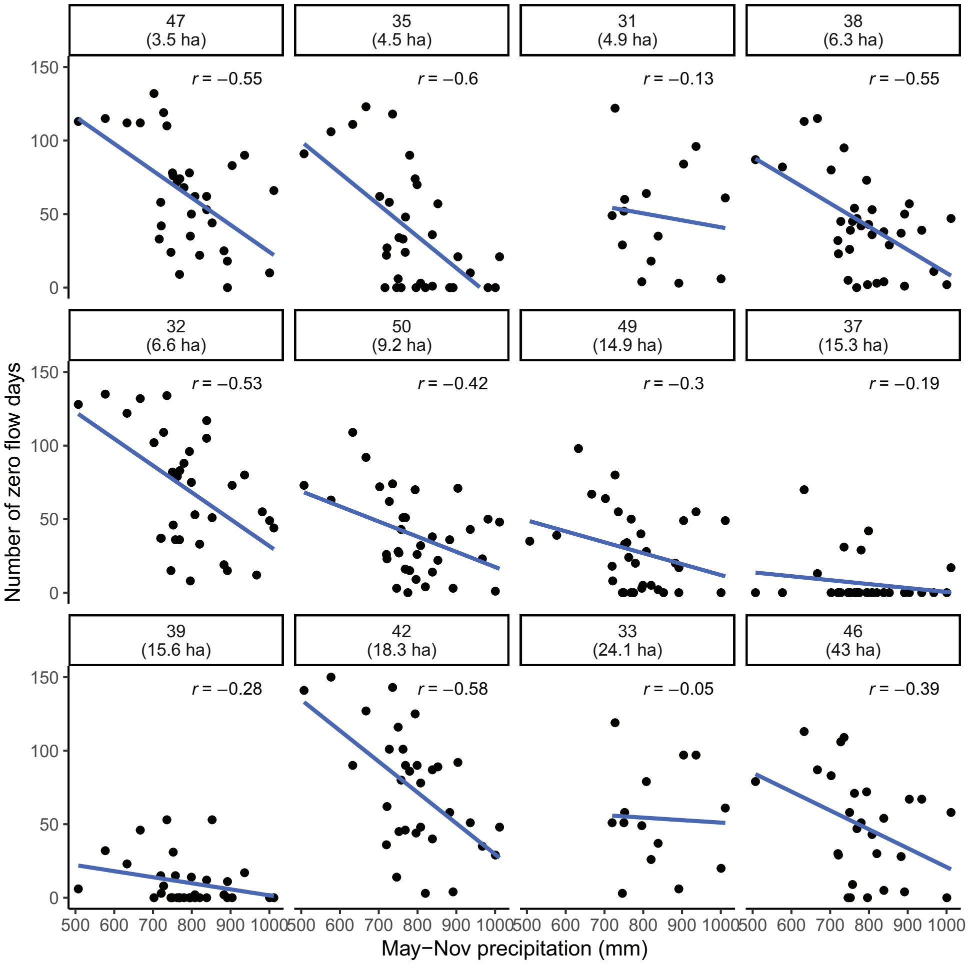

FIGURE 7 | Number of zero-flow days plotted against May–November precipitation for the 13 streams. Line of best fit is shown in blue and the Pearson correlations (r) are shown inset. Only pre-harvest years (1983–1996) are shown for the two catchments that were harvested (c31 and c33). Note that c34 is omitted as it did not have any zero-flow days during the pre-harvest period. Years with more than 20% missing flow data are not shown. Individual plots are labelled with the catchment number and the catchment area in parentheses. Plots are ordered by catchment area from top-left to bottom-right.

4 | Discussion

Temporal variability in number of zero-flow days was more strongly related to May–November meteorology than to antecedent snow conditions despite the large snowpacks that develop at this site. Different meteorological metrics representing the influence of precipitation and evapotranspiration have been consistently associated with flow permanence across various hydroclimatic environments, with drier and warmer years or regions having greater number of zero-flow days than wetter and colder years or regions (e.g., Reynolds, Shafroth, and Poff 2015; Eng, Wolock, and Dettinger 2016; Costigan et al. 2016; Jaeger et al. 2019; Sauquet et al. 2021a; Tramblay et al. 2021). This is not surprising given that precipitation inputs and evapotranspiration losses are first order controls on catchment water balances (Buttle et al. 2012). Most research on temporary streams is from arid and Mediterranean climates with extended periods of minimal precipitation during summer (e.g., Jaeger, Olden, and Pelland 2014; Pate, Segura, and Bladon 2020). In contrast, our study site has a continental climate with relatively even precipitation throughout the year (Figure 3). Despite these precipitation inputs, many of these streams still have extended dry periods except during the wettest years, likely reflecting the strong influence of transpiration losses on streamflow generation during the summer months in this forested environment.

Studies, primarily from mountainous environments, have found strong linkages between winter snow accumulation and summer low flow and zero-flow conditions (Godsey and Kirchner 2014; Sando and Blasch 2015; Jaeger et al. 2019). We found weak relationships between antecedent snow and the number of zero-flow days despite substantial variability in inter-annual snow conditions (Figure 8). These contrasting findings may be due to snowpacks in high elevation mountainous environments persisting much longer through spring and into summer than what occurs at the Turkey Lakes watershed, where snow typically disappears in April and early May. The later snowmelt at high elevations can provide a key source of water that sustains surface flow into the summer. In addition, Turkey Lakes is located on the Canadian Shield, which is characterised by shallow soils overlaying relatively impermeable till and bedrock; therefore, there may be limited subsurface storage capacity during snowmelt (Bottomley, Craig, and Johnston 1986). Regardless of the variability in snowpack size between years, any differences

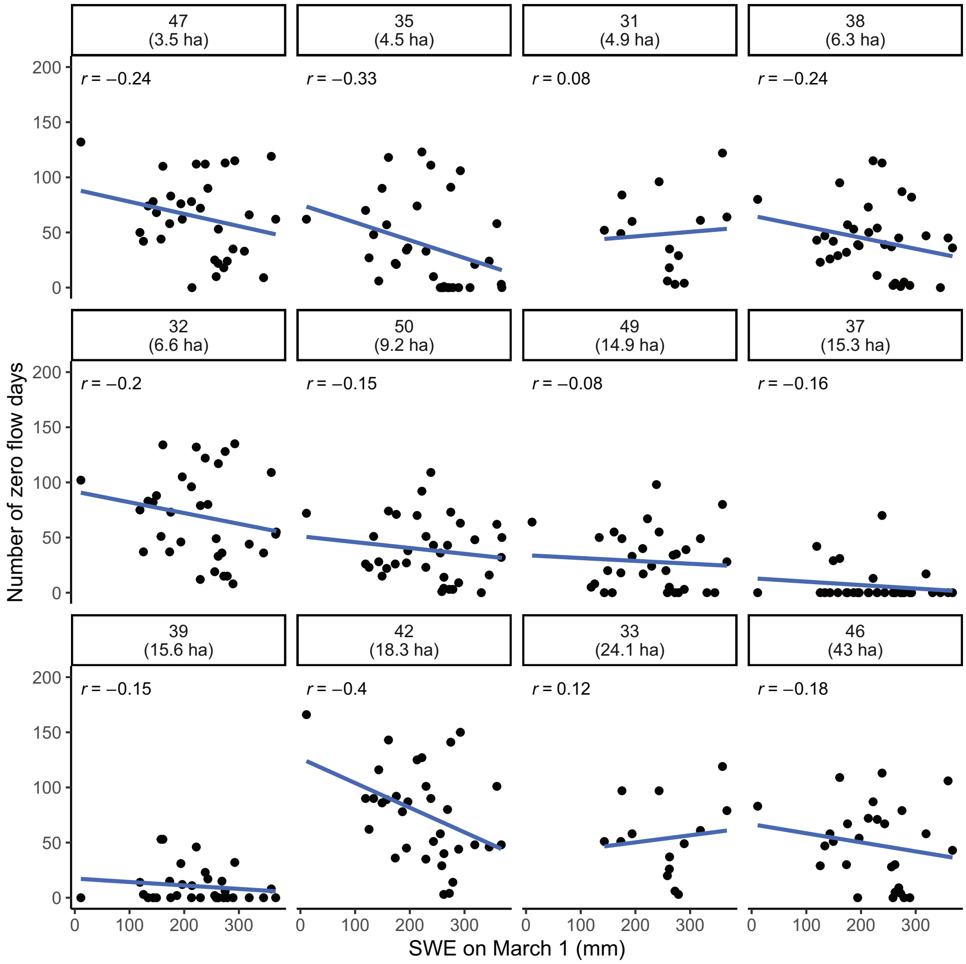

FIGURE 8 | Number of zero-flow days plotted against 1 March SWE for the 13 streams. Line of best fit is shown in blue and the Pearson correlations (r) are shown inset. Only pre-harvest years (1983–1996) are shown for the two catchments that were harvested (c31 and c33). Note that c34 is omitted as it did not have any zero-flow days during the pre-harvest period. Years with more than 20% missing flow data are not shown. Individual plots are labelled with the catchment number and the catchment area in parentheses. Plots are ordered by catchment area from top-left to bottom-right.

in antecedent conditions are likely reset during freshet, resulting in the main driver of inter-annual variability in zero-flow days being the meteorological conditions of the current year. In contrast, catchments with deeper soils may be able to store more snowmelt, which can then contribute to flow later in the summer (Godsey and Kirchner 2014). It is also possible that the weak relationship between snow and number of zero-flow days is due to uncertainty in the modelled snow estimates. Although the snow accumulation and melt model generally captured historic patterns in measured SWE (root mean square error of ±44 mm), there were some years where the model over- or under-estimated observed SWE by up to 100 mm (Leach et al. 2020).

Despite the streams being co-located within a small geographical area and subject to similar climate, they exhibited substantial variability in the number of zero-flow days for a given year. Based on previous studies (e.g., Svec, Kolka, and Stringer 2005; Pate, Segura, and Bladon 2020; Sando et al. 2022), we expected to find larger catchments associated with greater flow permanence, which was generally consistent with our results. However, that relationship was strongly influenced by the largest catchment in our study, c34, which did not have a documented zero-flow day during 1983–1996 (prior to being harvested in 1997), although the catchment did have a few zero-flow days during the drier post-harvest period (8 days in 2004 and 5 days in 2005). Ignoring c34, the relationship between catchment area and number of zero-flow days was weaker, as some of the streams with similar catchment areas had strongly contrasting numbers of zero-flow days (e.g., c39 and c42; Figure 5), suggesting other factors may be influencing spatial variability in flow duration.

Catchment characteristics that are often associated with slower flowpaths to the stream, such as low mean slope angles, large mean TWI values, presence of wetlands and slower mean travel times, are often found to be correlated with greater flow duration (Svec, Kolka, and Stringer 2005; Jensen et al. 2018; Prancevic and Kirchner 2019; Sando et al. 2022; Barua et al. 2022). We found generally similar, albeit noisy, relationships for our study catchments (Figures 10 and 11). However, not all catchment characteristics exhibited expected relationships with number of zero-flow days. We found almost no correlation between number of zero-flow days and mean travel time or mean flowpath length. A possible explanation for a lack of relationship could be that these metrics are inadequate proxies for hydrologic controls

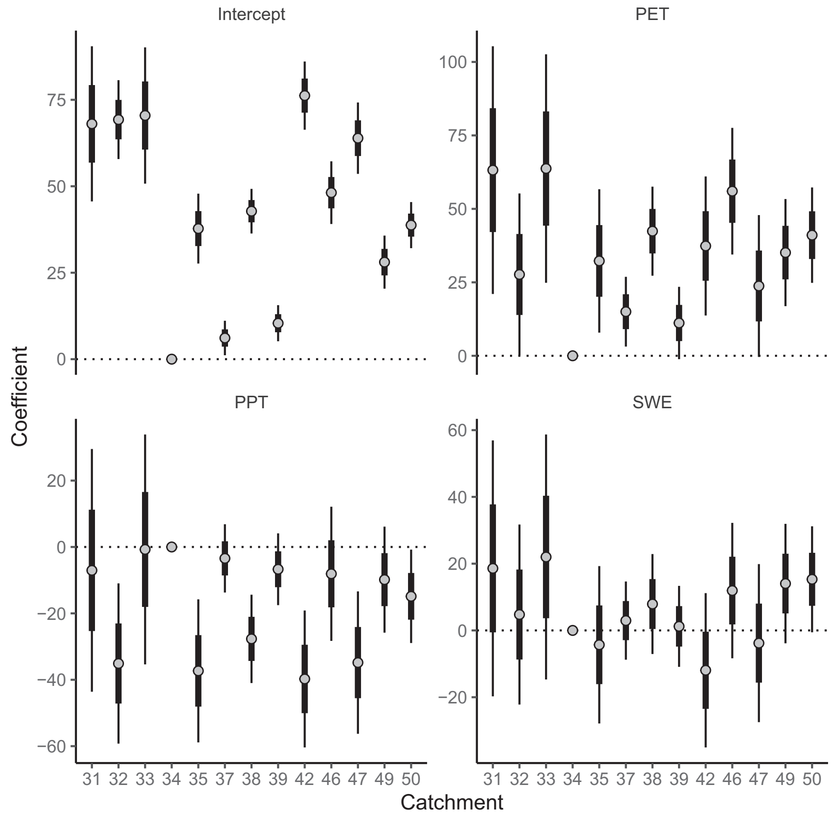

FIGURE 9 | Coefficient estimates from the linear regression model fit to each stream. Predictor variables have been scaled and standardised. Intercepts represent the mean number of zero-flow days. The thick line represents the 50% uncertainty interval and the thin line represents the 95% uncertainty interval.

on stream drying in our system. Leach et al. (2020) found mean travel time estimates for the Turkey Lakes catchments and their relationships to catchment characteristics changed between wet and dry periods. For example, wetland cover was associated with shorter mean travel times during wet periods but longer mean travel times during dry periods; therefore, suggesting that longer mean travel times are not a direct indicator of water availability to sustain surface flow. In addition, there is large uncertainty in mean travel time estimation, which may confound potential empirical relationships (Leach et al. 2020).

We also used winter stream temperature as a proxy for groundwater influence and found a negative relationship between winter stream temperature and number of zero-flow days. This is consistent with the expectation that catchments with greater groundwater influence should exhibit greater flow duration. Warix et al. (2021) investigated relationships between patterns of stream drying and topographic and groundwater influences for a semi-arid catchment in Idaho. They found that some metrics associated with longer water residence times and greater subsurface storage tended to be related to greater flow permanence; however, similar to our study, these variables only explained a limited amount (< 30%) of the variability in observed flow permanence.

Overall, we were less successful at accounting for spatial variability than temporal variability in number of zero-flow days. This could be due to the predictor variables considered in this study being imperfect proxies for actual water flowpaths and storage mechanisms occurring in these catchments. It is also possible that we missed representing important processes such as hyporheic storage and its influence on localised stream drying (Sando and Blasch 2015). It is likely that use of hydrometric station observations is better at capturing temporal variability than spatial variability in zero-flow occurrences. Zero-flow observations can be uncertain due to measurement challenges (Zimmer et al. 2020). Leakage beneath the weir or factors controlling localised drying around the weir are not included as predictor variables in the spatial analysis. These factors can confound spatial comparisons, but temporal comparisons may be more robust since these processes may influence stream drying in a similar manner for a given site from year-to-year. Hydrometric observations are relevant for the stream reach where the station is located and may not be representative of conditions up and downstream of the monitored reach. Field observations by staff working at the site have noted that when the streambed is dry at the weir, the entire upstream network is also usually dry for many streams; however, some streams can have isolated small pools upstream of the weir when the weir is dry. Surveys of surface water presence along the channel networks, combined with the hydrometric station observations, could provide more insight into flow duration dynamics.

Results from our study suggest that zero-flow occurrences at Turkey Lakes may be more sensitive to future changes in

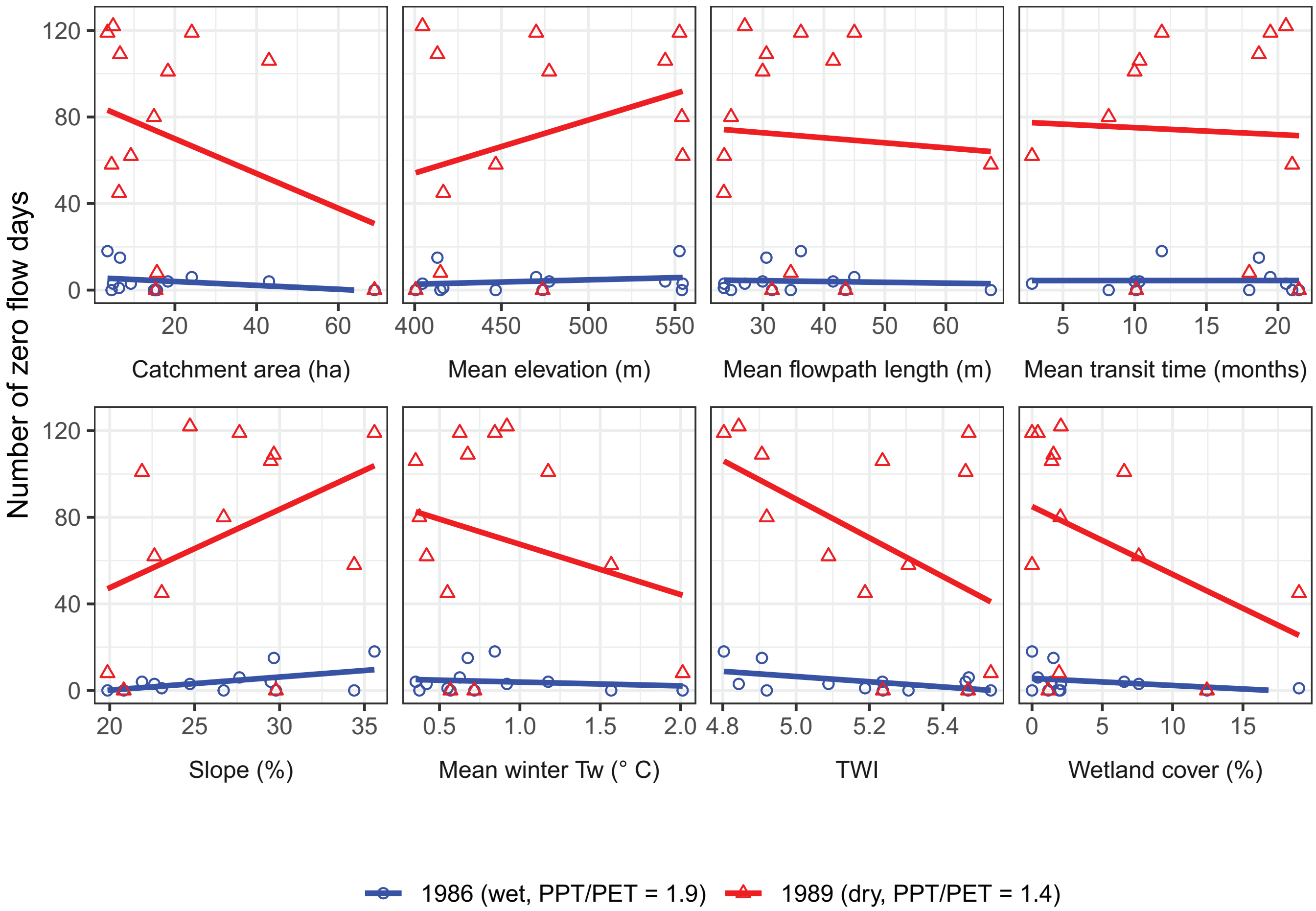

FIGURE 10 | Number of zero flow days plotted against various catchment characteristics (catchment area, mean flowpath length, mean transit time, mean winter stream temperature, mean slope, mean elevation, mean topographic wetness index (TWI) and percent wetland cover). Values from a relatively wet year (1986) are shown in blue and values from a relatively dry year (1989) are shown in red. A line of best fit is included to help with visualisation.

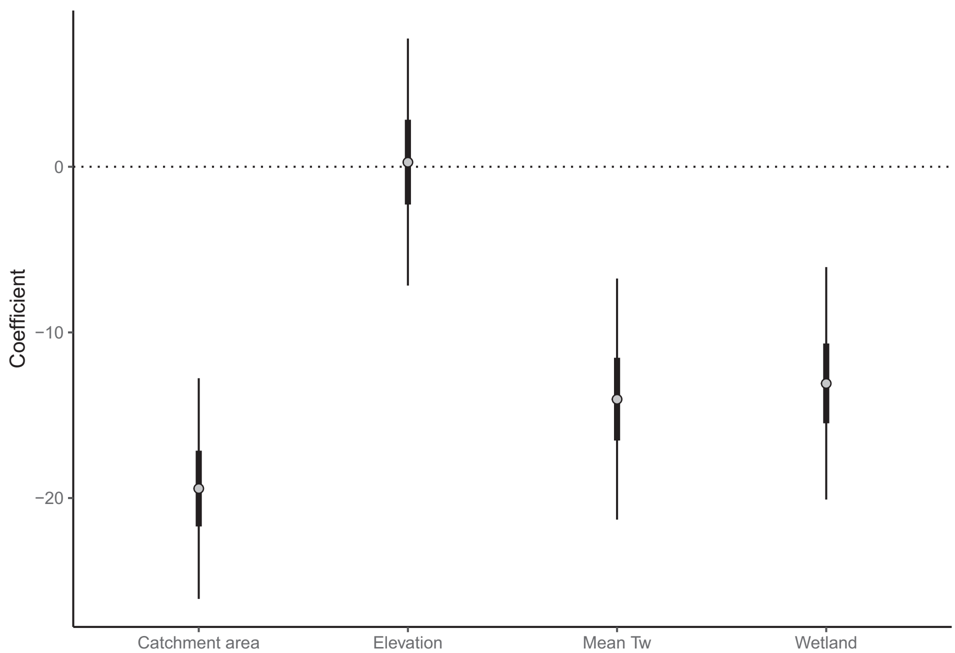

FIGURE 11 | Coefficient estimates from the linear mixed model fit to spatial variability in zero-flow days. Coefficient estimates are for catchment area, mean catchment elevation, mean winter stream temperature (Mean Tw) and percent wetland cover. Predictor variables were scaled and standardised. The thick line represents the 50% uncertainty interval and the thin line represents the 95% uncertainty interval.

weather conditions during the spring to fall period compared to potential changes in snow. Climate change projections for this region of Ontario suggest a future with higher air temperatures and more precipitation (Wang et al. 2015). These projected changes in air temperature and precipitation may have counteracting effects on flow duration of these headwater streams. Higher air temperatures may be associated with elevated evapotranspiration rates, increasing the number of zero-flow days. In contrast, more precipitation could contribute to lower frequencies of zero-flow days. Given our empirical analysis, it is difficult to disentangle the relative influence of precipitation and potential evapotranspiration on the number of zero-flow days. Potential evapotranspiration had better explanatory power than precipitation when accounting for variability in number of zero-flow days, suggesting the projected increases in air temperature may be more influential than the increases in precipitation. However, more process-based investigations on the drivers of stream drying are needed in order to make rigorous predictions about how streams will respond to future changes in climate.

5 | Conclusion

We show that forested headwater streams in humid environments can lack surface flow for extended periods and that these zero-flow dynamics are tightly coupled to meteorological conditions. Despite relatively even precipitation inputs throughout the year, many streams dry up during the summer. Spring to fall meteorological conditions, in particular potential evapotranspiration, explained between 18% and 76% of the variability in number of zero-flow days, depending on catchment. In contrast, antecedent snow conditions explained little (< 16%) of the temporal variability in zero-flow days. Catchment characteristics often associated with slower water delivery from hillslopes to streams, as well as indicators of greater groundwater influence, tended to be negatively correlated with number of zero-flow days across the sites. However, our spatial comparisons were noisy and more research is needed to understand how catchment structure may moderate or amplify climate change effects on stream drying.

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Data Availability Statement

The data that support the findings of this study are openly available in Zenodo at https://zenodo.org/records/10961029.

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