Received: 15 May 2017 | Accepted: 27 November 2017
DOI: 10.1002/hyp.11418

RESEARCH ARTICLE

WILEY

Mediating stream baseflow response to climate change: The role of basin storage

James M. Buttle

School of the Environment, Trent University, Peterborough, ON K9L 0G2, Canada

Correspondence
James M. Buttle, School of the Environment, Trent University, Peterborough, ON K9L 0G2, Canada.
Email: jbuttle@trentu.ca

Funding information
Natural Sciences and Engineering Research Council of Canada, Grant/Award Number: RGPIN-2015-06116

Abstract

Inter-basin differences in streamflow response to changes in regional hydroclimatology may reflect variations in storage characteristics that control the retention and release of water inputs. These aspects of storage could mediate a basin's sensitivity to climate change. The hypothesis that temporal trends in stream baseflow exhibit a more muted reaction to changes in precipitation and evapotranspiration for basins with greater storage was tested on the Oak Ridges Moraine (ORM) in Southern Ontario, Canada. Long-term (>25 years) baseflow trends for 16 basins were compared to corresponding trends in precipitation amount and type and in potential evapotranspiration as well as shorter trends in groundwater levels for monitoring wells on the ORM. Inter-basin differences in storage properties were characterized using physiographic, hydrogeologic, land use/land cover, and streamflow metrics. The latter included the slope of the basin's flow duration curve and basin dynamic storage. Most basins showed temporal increases in baseflow, consistent with limited evidence of increases and decreases in regional precipitation and snowfall: precipitation ratio, respectively, and recent increases in groundwater recharge along the crest of the ORM. Baseflow trend magnitude was uncorrelated to basin physiographic, hydrogeologic, land use/land cover, or flow duration curve characteristics. However, it was positively related to a basin's dynamic storage, particularly for basins with limited coverage of open water and wetlands. The dynamic storage approach assumes that a basin behaves as a first-order dynamical system, and extensive open water and wetland areas in a basin may invalidate this assumption. Previous work suggested that smaller dynamic storage was linked to greater damping of temporal variations in water inputs and reduced interannual variability in streamflow regime. Storage and release of water inputs to a basin may assist in mediating baseflow response to temporal changes in regional hydroclimatology and may partly account for inter-basin differences in that response. Such storage characteristics should be considered when forecasting the impacts of climate change on regional streamflow.

KEYWORDS

baseflow, climate change, dynamic storage, flow duration curve, hydrogeology, land use/land cover, oak ridges moraine, physiography, storage

1 | INTRODUCTION

Estimating streamflow response to a changing climate is a major challenge facing hydrology (Singh, Wagener, Werkhoven, Mann, & Crane, 2011). Forecasts of the impacts of climate change on streamflow have been made at the global, national, regional, and basin scales. For example, Barnett, Adam, and Lettenmaier (2005) used a spatially distributed macroscale hydrology model to assess the importance of snow to annual run-off at the global scale and how the reaction of snow cover to global warming may influence run-off. Berghuijs, Woods, and Hrachowitz (2014) applied the Budyko water balance framework to basins in the contiguous United States where they found that basins with a higher fraction of annual precipitation falling as snow had higher mean streamflow compared with those with marginal or no snowfall, implying a decrease in the snowfall : precipitation ratio will result in decreased streamflow. Regional scale examples include Hayhoe et al.'s (2007) use of atmosphere–ocean general circulation models to reproduce 1970–2000 temporal trends in observed climatological and hydrological indicators in the U.S. north-east, which included increases in annual temperature and declines in annual precipitation, snowfall, evaporation, run-off, and low flows. They forecast increases in temperature, annual, and winter precipitation for the 2070–2090 period, but little change in summer precipitation. This was predicted to lead to increases in annual evaporation and run-off but continued declines in snowfall and stream low flows. Cherkauer and Sinha (2010) projected increased winter and spring low flows in the Lake Michigan region as a result of greater precipitation and warmer temperatures, which reduce snow accumulation and soil frost development. However, summer low flows were most likely to decline due to higher evapotranspiration and possible decreases in summer precipitation. At the basin scale, Sultana and Coulibaly (2011) used downscaled global climate model data from Southern Ontario as input to the coupled MIKE SHE/MIKE 11 hydrologic model to predict hydrologic response to forecast increases in precipitation and air temperature for a small basin west of Toronto, Ontario. They suggested a 1–10% increase in evapotranspiration by the 2046–2065 period would lead to reductions in groundwater recharge and stream baseflow, despite an anticipated 14–17% increase in annual mean precipitation over the same time interval.

These projections may be of great value to water resource managers operating in the face of a changing climate. However, the question of if these projected changes in streamflow reflect existing relationships between trends in climate and streamflow metrics often remains unanswered. Whether current relationships between hydroclimatic drivers and resultant streamflow will persist in the future is unclear; nevertheless, such relationships and the physical insights underpinning them may assist in assessing the validity of forecasts of streamflow response to climate change. Our current knowledge of how climate change will affect stream baseflow (*BF*) is particularly limited (Price, 2011). Baseflow is an important component of basin streamflow, because it provides an estimate of groundwater recharge at the basin scale, is a vehicle for a variety of aquatic ecosystem services (e.g., maintaining fish habitat), and is an index of water availability for human uses such as extraction of water for irrigation and dilution of wastewater effluents (Cherkauer & Sinha, 2010). Rivard, Vigneault, Piggott, Larocque, and Anctil (2009) note that *BF* time series can serve as a proxy of trends of recharge and may provide insight into how groundwater recharge reacts to climate change.

Inter-regional differences in stream *BF* response to climate change will likely be largely driven by shifts in the relative magnitude of precipitation and evapotranspiration (Ficklin, Robeson, & Knouft, 2016); however, not all basins within a given region facing relatively uniform climatic changes will necessarily experience a similar *BF* response (Cheng, Li, Li, & Auld, 2012; Ficklin et al., 2016). Thus, Kling et al. (2003) showed great variability in stream reactions to climate-driven changes across the Great Lakes region, largely resulting from differences in the relative contribution of groundwater versus surface water to their flow patterns. They suggested that wetland systems (and by extension stream channel networks) that are largely recharged by groundwater are more resistant to climate-driven changes. Knowledge of the hydrologic function of key basin properties may therefore contribute to understanding how *BF* will respond to climate change. One such property is basin storage (Soulsby, Piegat, Seibert, & Tetzlaff, 2011), whose role is linked to the concept of hydrologic resistance: the degree of desynchronization between a basin's precipitation inputs and streamflow outputs (Carey et al., 2010). Basins with greater hydrologic resistance would store water over long time periods and gradually release it as streamflow. Such basins may exhibit a more muted sensitivity to temporal changes in hydroclimatic drivers such as precipitation and evapotranspiration compared to those with smaller hydrologic resistance and greater interannual synchrony between inputs and outputs. This is supported by Cooper, Wilkinson, and Arnell's (1995) modelling study of the effects of climate change on aquifer storage and stream *BF* for sandstone and chalk aquifers in the UK. The order-of-magnitude larger storage coefficient for the sandstone aquifer led to less severe reductions in simulated *BF* following climate change relative to the chalk aquifer. Storage in the former also showed a longer time needed to reach equilibrium in response to a given change in climate.

The Oak Ridges Moraine (ORM) is an important hydrogeologic feature in Southern Ontario (Holysh & Gerber, 2014), supplying potable water to ~60,000 wells (Sharpe, Russell, & Logan, 2007) and more than 250,000 people (Furberg & Ban, 2012). Studies of the linkage between climate change and stream *BF* response in the ORM region benefit from the numerous streams that are gauged in the region, the relatively high density of climate stations for assessing temporal trends in key hydroclimatic drivers and the detailed hydrogeologic information available for the region. Buttle, Greenwood, and Gerber (2015) suggested that *BF* sensitivity to an increase in annual precipitation over the last several decades for streams draining the ORM has been non-uniform, with some basins showing increased *BF* with time, whereas others showed no significant temporal trends. However, streamflow records used in this analysis were not all for the same time period, and some were as short as 13 years. Therefore, the purpose of this paper is fourfold:

1. To examine longer term (>25 years) concurrent temporal trends in *BF* for streams draining the ORM;
2. To compare these trends to those in such hydroclimatic variables as precipitation and potential evapotranspiration as well as trends in groundwater levels for monitoring wells on the ORM;
3. To assess whether inter-basin differences in *BF* trends can be explained by metrics of basin physiography, hydrogeology, land use/land cover (LULC) and streamflow that could be related to basin storage; and
4. To test the hypothesis that temporal trends in *BF* for basins with greater storage potential exhibit a more muted response to recent trends in the hydroclimatology of the ORM region relative to basins with smaller storage potential.

2 | STUDY AREA AND METHODS

2.1 | Study area

The ORM (Figure 1) is an interlobate kame moraine, the crest of which consists of permeable Quaternary glaciofluvial gravel and sand ice-

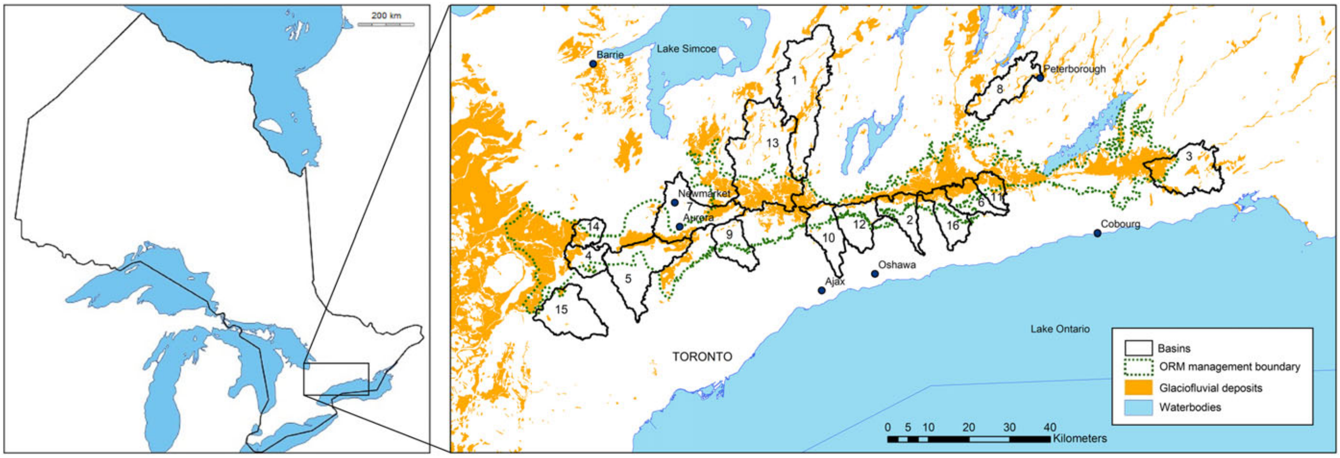

FIGURE 1 Selected basins draining the Oak Ridges Moraine (ORM) with >25 years of streamflow records to 2012 from the Water Survey of Canada, the ORM management boundary, and the extent of outcropping glaciofluvial deposits (OGS, 2003). Characteristics of each of the numbered basins are given in Table 1

contact deposits (the Oak Ridges Moraine Aquifer Complex, ORMAC) > 100 m thick in places (Gerber & Howard, 2000, 2002). Less permeable clayey silt-to-silt and sandy silt-to-sand tills overlie and extend beyond the ORMAC on the ORM's north and south flanks. Thus, the ORMAC is an unconfined/partly confined aquifer that rests in turn on alternating Quaternary aquitards and aquifers overlying Ordovician shale, limestone, dolostone, and siltstone bedrock. Soils are brunisolic grey brown luvisols (Soil Classification Working Group, 1998; FAO equivalent: arenosol), with sand or sandy loam textures along the crest of the ORM and sandy loams and loams on till units (Buttle, 2011). Soils on the crest of the ORM have large saturated hydraulic conductivities with mean values to 1 m depth exceeding 260 mm h^{-1} (Greenwood & Buttle, 2014), whereas the flanking tills are much less permeable (Gerber & Howard, 2000). More detailed descriptions of the ORM's hydrogeology are given in Gerber and Howard (2002) and Sharpe, Hinton, Russell, and Desbarats (2002), whereas Buttle et al. (2015) summarize the ORM's climatology, physiography, hydrology, and land cover.

2.2 | METHODS

2.2.1 | Hydroclimatic variables

Precipitation records for 11 Meteorological Service of Canada (MSC) climate stations in the ORM region were examined (data accessed at http://climate.weather.gc.ca/historical_data/search_historic_data_e.html). Stations were selected on the basis of (a) length of record, (b) geographical coverage of the region, and (c) most recent data in order to relate temporal trends in precipitation to streamflow data which extend to 2012. Climate record lengths ranged from 26 years (one MSC station) to 63 years (two MSC stations), with an average record length of 45 years (±12 years *SD*). The following values were abstracted: (a) water year (WY, October 1–September 30) total precipitation (*P*), WY snowfall, and WY snow: *P*; (b) *P*, snow, and snow: *P* for winter (DJF), spring (MAM), summer (JJA), and fall (SON); and (c) winter *P*:WY *P*, spring *P*:WY *P*, and winter + spring *P*:WY *P*. The temperature-based Hamon model (Dingman, 2002) was used to estimate annual potential evapotranspiration (*PET*) for all MSC stations (eight) with air temperature data extending into the 2000s.

2.2.2 | Groundwater levels

Water level records were obtained from Ontario Provincial Groundwater Monitoring Network wells located within the ORM planning boundary, defined as the 245 m asl topographic contour (Gerber & Howard, 2002), and screened within the sand and gravel deposits of the ORMAC (data accessed at https://www.ontario.ca/environment-and-energy/map-provincial-groundwater-monitoring-network). Data from 19 wells with records spanning the period from 2001 to 2012 were obtained (Figure 2), with record lengths ranging from 3 (one well) to 11 years (five wells). Time series of monthly average water levels for each well were derived from continuously measured water levels recorded at hourly intervals.

2.2.3 | Baseflow

Basins gauged by the Water Survey of Canada (WSC) with headwaters within the ORM planning boundary and with at least 25 years of record extending to 2012 were examined, with the exception of WSC stations on lower reaches of the Humber and Don Rivers. These rivers flow through Toronto, and increased stormflow run-off in this highly urbanized landscape might distort their BF estimates. Rivers with significant flow regulation, as defined by the WSC and Moin and Shaw (1985), were removed because flow regulation may result in greater than-normal low-flow characteristics due to such issues as required flow releases from dams (Stuckey, 2006). Unregulated flow records are also preferred for assessing climate–streamflow relationships (Ficklin et al., 2016). This resulted in records (data accessed from https://wateroffice.ec.gc.ca/search/historical_e.html) from 16 WSC stations (Figure 1). The Web-Based Hydrological Analysis Tool (Lim et al., 2005) was used to separate *BF* from total streamflow using the one parameter digital filter, with the filter parameter (0.925) held constant for the entire period of record. Daily *BF* was converted to mm day^{-1} using basin areas and summed to give WY *BF* depths. These are applied to the entire basin area upstream of the WSC gauge, and only complete WY data were used.

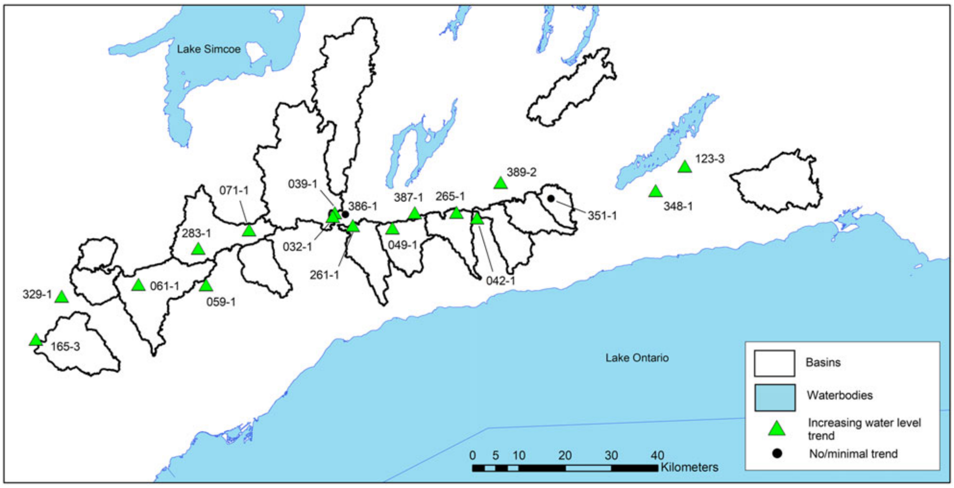

FIGURE 2 Location of Ontario provincial groundwater monitoring wells within the ORM planning boundary and screened within the sand and gravel deposits of the Oak Ridges Moraine aquifer complex. Well identification number prefix “W0000” has been removed for all wells to improve clarity. Well water level trends are presented in Figure 7

2.2.4 | Temporal trends

Monotonic temporal trends in hydroclimatic and *BF* data were analysed using Mann–Kendall tests, adjusted for temporal autocorrelation in the data (Clarke, Hulley, Marsalek, & Watt, 2011). Equations for trend lines were estimated using the Kendall–Theil robust line (Version 1.0, Granato, 2006), a nonparametric regression approach resistant to outliers and skewness (Meyer, 2005). Temporal trends significant at the p = .05 level are reported.

2.2.5 | Streamflow-based metrics of basin storage

Daily streamflows for the 2006–2012 WYs for all basins were used to derive flow duration curves. The Q_5, Q_{50}, and Q_{95} flow percentiles were abstracted, and two metrics were derived: Q_5–Q_{95} and $(Q_5–Q_{95})/Q_{50}$. Both indices provide a measure of the slope of the flow duration curve, where a large value indicates a variable flow regime, and a small value means a more damped response and relatively high storage within the basin (Sawicz, Wagener, Sivapalan, Troch, & Carrillo, 2011). Dynamic storage is that portion of basin storage that is hydrologically active and contributes directly to streamflow (McNamara et al., 2011). Dynamic storage was estimated for the 2006–2012 WYs for each basin following Kirchner's (2009) method of assessing the sensitivity of basin discharge Q (mm day^{-1}) to changes in storage S (mm):

$$\frac{dQ}{dt} = \frac{dQ}{dS}\frac{dS}{dt} = g(Q)(P\text{–}ET\text{–}Q) \tag{1}$$

where P is precipitation (mm day^{-1}), ET is evapotranspiration (mm day^{-1}), $g(Q)$ is the discharge sensitivity function, and there is no water loss to deep groundwater recharge. The $g(Q)$ function can be estimated during streamflow recessions when P and ET are negligible as

$$g(Q) = \frac{dQ}{dS} \approx \left.\frac{-\frac{dQ}{dt}}{Q}\right|_{P \ll Q, ET \ll Q} \tag{2}$$

The S–Q relationship can be derived as

$$\int dS = \int \frac{dQ}{g(Q)} \tag{3}$$

which can be inverted to obtain Q as a function of S. Kirchner's (2009) approach was modified because Q was available as mean daily rather than hourly discharge. In order to maximize the number of days with streamflow data that could be used to derive $g(Q)$ while attempting to ensure that flow recessions were not affected by P or ET, recessions were examined for periods with no P (from the nearest MSC climate station), no or limited ET (November to April, a period when daily air temperatures across the ORM region are often below 0°C and the region is frequently snow-covered) and no snowmelt input (using air temperatures at the nearest MSC climate station). Kirchner's (2009) data binning procedure was followed to derive the $g(Q)$ function from second-order polynomial regression of ln(−dQ/dt) on ln$Q$, adjusted for the smaller number of observations arising from daily rather than hourly discharges. Kirchner (2009) defined dynamic storage as the difference in S between wet and dry periods in the basin, which he assumed to be represented by the difference between maximum and minimum S at the average annual maximum and minimum Q for the period of record. Dynamic storage values were obtained by expressing the S values obtained using equation 3 relative to an assumed S of 0 mm at mean Q for the 2006–2012 WYs (see Buttle, 2016 for an example derivation of dynamic storage for a basin draining the ORM). As Teuling, Lehner, Kirchner, and Seneviratne (2010) noted regarding the Kirchner (2009) approach to estimating dynamic storage: “The only (but necessary) assumption is that runoff is solely dependent on the total water storage in the catchment” (p 1). This assumption was met by deriving the $g(Q)$ function for the study basins for periods with minimal P and ET, as noted earlier.

2.2.6 | Basin physiographic metrics

WSC gauging station locations were plotted on the Ontario provincial 10-m resolution digital elevation model (DEM, OMNR, 2006) in

ArcGIS10, and basin boundaries and area for each station were determined using the "Hydrology" toolbox. Mean basin slope was estimated as the arithmetic mean slope of all DEM 10 × 10-m cells in the basin (McLean & Watt, 2005). Streamflow from basins with steeper slopes might receive greater contributions from relatively rapid run-off flowpaths (Price, 2011), thus reducing the opportunity for storage of precipitation inputs in the basin's unsaturated and saturated zones.

2.2.7 | Land use/land cover metrics

Spatial extent of major LULC types in each basin for the 2000–2002 period was estimated using the Southern Ontario Land Resource Information System (SOLRIS) of the Ontario Ministry of Natural Resources (SOLRIS Technical team, 2008). Five distinct LULC types were abstracted: agricultural (both intensive, such as row cropping, and nonintensive, such as pasture), forest, urban, extraction (pits and quarries), and open water plus wetlands. The fractional coverage of each LULC type was determined for each basin. Forest land cover on the ORM promotes greater infiltration and subsurface storage of precipitation inputs relative to agricultural (Greenwood & Buttle, 2014) and urban LULCs (Hubbart & Zell, 2013).

2.2.8 | Hydrogeologic metrics

The extent of outcropping glaciofluvial deposits (GFD) consisting of stratified ice contact, outwash, and glaciodeltaic deposits (predominantly sands and gravels) was obtained from the Ontario Geological Survey's surficial geology digital coverage (OGS, 2003) and imported into ArcGIS10. These GFD are the outcrop of the ORMAC that underlies the crest of the ORM, and the fractions of each basin underlain by GFD were determined. A detailed three-dimensional regional hydrogeologic model of the ORM (Holysh & Gerber, 2014) was used to obtain several metrics of storage-related properties of the ORMAC. The mean static level surface (assumed to equal the water table position) within the ORMAC estimated from 1,000s of water well records in the region was subtracted from surface elevations to derive the mean and standard deviation of depth to water table (WT) for each basin to characterize its unsaturated zone. The mean and standard deviation of the thickness of the ORMAC for each basin were also obtained. It was assumed that storage potential would be maximized in basins with extensive outcropping GFD and deep unsaturated zones and ORMAC deposits of relatively uniform thickness.

2.2.9 | Statistical analyses

Untransformed and transformed (logarithmic, square-root, cube-root) forms of all variables were tested for normality using the Shapiro–Wilks test. Correlations (Pearson's r) between variables were determined. Many variables exhibited collinearity (e.g., mean basin slope and % GFD); therefore, principal components analysis (PCA) was used to identify groups of correlated and normalized basin characteristics for use as independent variables. PCA was applied with no rotation to the correlation matrix of basin characteristics and all principal components (PCs) with eigenvalues greater than 1 were retained. Flow duration curve and dynamic storage metrics were regressed on PC scores to determine if inter-basin differences in these streamflow-based storage metrics could be explained by basin characteristics. Similarly, annual trends in BF were regressed on PC scores as well as on the streamflow-based storage metrics.

3 | RESULTS

3.1 | Basin physiography and land cover

All basins contained outcropping ORMAC deposits, ranging from 1.4% to 41.7% of the basin area (Table 1). Mean basin slope ranged from 2.8% to 7.9% and was significantly correlated with % coverage of outcropping ORMAC (r = .75, p < .05). Mean thickness of the ORMAC ranged from as little as 1 m to as much as 54 m. Basins had a considerable unsaturated zone, with an average mean depth to WT of 11 m. Mean depth to WT varied by a factor of 5 between basins, and ln(mean depth to WT) was significantly correlated with mean basin slope (r = .91, p < .05). Agriculture was the dominant land cover (mean coverage of 60.1 ± 12.6%) and was inversely correlated with % forest cover (r = −.77, p < .05). All basins had forest cover in their headwaters and stream valleys, whereas excavation areas (gravel pits and quarries) were generally insignificant. Most basins had less than 10% urban cover, although 02EC009—Holland R at Holland Landing had >30% urban area. All basins had some open water and wetland cover; however, this was generally minor with the exception of larger basins on the ORM's northern and eastern flank (02EC011—Beaverton R at Beaverton, 02HJ001—Jackson Cr at Peterborough, 02EC018—Pefferlaw Br nr Udora, 02HK007—Cold Cr at Orland).

3.2 | Streamflow-based storage metrics

Slopes of the flow duration curves for the 2006–2012 WYs as represented by Q_5–Q_{95} ranged from 1.5 to 3.6 mm day^{-1} (Table 2, Figure 3). Standardizing these slopes by Q_{50} increased the range in values, and there was a significant correlation between (Q_5–Q_{95})/Q_{50} and Q_5–Q_{95} (r = .55, p < .05).

Estimation of dynamic storage using Equations 1–3 requires streamflow recession data that are not influenced by either P or ET. Whereas the restriction of daily streamflows used to derive $g(Q)$ to periods with no P or snowmelt inputs is straight-forward, the issue of the potential influence of ET on daily streamflows is more problematic. It was assumed that this influence would be minimized by using streamflow recessions for the November to April period. Evapotranspiration begins to increase in April following negligible ET from November to March in the ORM region, although mean daily ET rates in April for typical surface types on the ORM are in the order of 1 mm day^{-1} or less (Delidjakova, Bello, & MacMillan, 2014). April streamflow recessions were included in the derivation of dynamic storage for many basins; however, they comprised an average of only 14% (±9%) of the daily flows used to derive $g(Q)$, and there was no relationship between dynamic storage and the fraction of April daily flows relative to the total number of daily flows used to estimate dynamic storage. Thus, any effect of ET on the derived dynamic storage values can be assumed to be minimal.

Dynamic storage ranged from 31 to 90 mm, and the range in values can be explained in part by inter-basin differences in hydrograph recession rates. Buttle (2016) demonstrated a decrease

TABLE 1 Study basin characteristics

Basin number in Figure 1; WSC station code; name	Area (km^2)	Mean slope (%)	% sand/gravel outcrop	Mean ORMAC[1] thickness (m)	CV[2] ORMAC thickness	Mean depth to WT[3] (m)	CV depth to WT	% agriculture	% forest	% urban	% extraction	% water + wetland
1; 02EC011; Beaverton R nr Beaverton	291.3	3.3	10.2	4	1.76	5	1.23	67.9	8.6	3.4	1.1	19.0
2; 02HD006; Bowmanville Cr at Bowmanville	82.9	6.0	19.1	22	0.95	11	0.95	622	22.8	6.3	0.2	8.5
3; 02HK007; Cold Cr at Orland	161.0	6.4	22.8	16	1.19	14	1.08	53.9	26.3	3.5	0.1	16.2
4; 02HC023; Cold Cr nr Bolton	62.8	6.8	18.2	54	0.51	10	0.78	67.6	15.6	9.0	0.1	7.7
5; 02HC009; East Humber R nr Pine Grove	190.6	5.5	12.0	43	0.70	11	0.87	56.2	19.1	17.7	0.1	6.8
6; 02HD003; Ganaraska R nr Osaca	75.3	7.7	34.0	39	0.91	26	0.93	41.2	47.0	2.6	0.3	8.9
7; 02EC009; Holland R at Holland Lndg	175.6	5.2	18.6	43	0.69	8	0.94	39.0	18.8	35.5	0.7	6.0
8; 02HJ001; Jackson Cr at Peterborough	110.0	5.2	7.7	1	1.55	8	1.29	56.5	10.2	9.4	0.2	23.8
9; 02HC028; Little Rouge Cr nr Locust Hill	82.7	2.8	1.4	13	0.93	5	1.25	75.8	9.4	8.1	0.3	6.4
10; 02HC018; Lynde Cr nr Whitby	100.3	4.7	7.8	16	1.14	7	1.06	63.6	17.3	12.3	0.6	6.3
11; 02HD004; NW Ganaraska nr Osaca	46.1	7.9	41.7	41	0.84	23	0.90	36.7	54.1	2.0	0.0	7.1
12; 02HD008; Oshawa Cr at Oshawa	95.8	5.7	3.6	15	0.95	8	0.86	74.2	13.7	7.8	0.2	4.1
13; 02EC018; Pefferlaw Br nr Udora	340.9	4.5	28.3	29	0.92	9	1.01	55.7	21.5	5.0	1.3	16.6
14; 02EC010; Schomberg R nr Schomberg	46.4	5.9	23.0	37	0.76	7	0.92	72.2	15.6	5.4	0.1	6.8
15; 02HC031; W Humber at Hwy 7	142.5	2.9	2.7	25	0.84	5	0.79	74.6	6.7	13.5	0.0	5.2
16; 02HD009; Wilmot Cr nr Newcastle	80.7	6.2	17.4	19	1.13	11	0.96	63.9	22.2	5.7	1.1	7.1
Mean	130.3	5.4	16.8	26	0.98	11	0.99	60.1	20.6	9.2	0.4	9.8
Standard deviation	85.0	1.5	11.5	15	0.32	6	0.16	12.6	13.0	8.2	0.4	5.8
CV	0.65	0.28	0.68	0.59	0.32	0.57	0.16	0.21	0.64	0.89	1.04	0.59
Maximum	340.9	7.9	41.7	54	1.76	26	1.29	75.8	54.1	35.5	1.3	23.8
Minimum	46.1	2.8	1.4	1	0.51	5	0.78	36.7	6.7	2.0	0.0	4.1

Note. WSC = Water Survey of Canada.

[1]ORMAC = Oak Ridges Moraine Aquifer Complex.

[2]CV = coefficient of variation.

[3]WT = water table.

TABLE 2 Flow duration curve slope, dynamic storage, and temporal trend in baseflow (*BF*, $p < .05$) for the study basins

Basin	Q_5–Q_{95} (mm day^{-1})	(Q_5–Q_{95})/Q_{50}	Dynamic storage (mm)	*BF* temporal trend (mm year^{-1})
02EC011; Beaverton R nr Beaverton	3.2	5.4	70	0.0
02HD006; Bowmanville Cr at Bowmanville	3.0	2.7	64	2.0
02HK007; Cold Cr at Orland	2.6	3.2	32	0.0
02HC023; Cold Cr nr Bolton	1.5	3.0	34	1.2
02HC009; East Humber R nr Pine Grove	2.0	4.5	33	0.9
02HD003; Ganaraska R nr Osaca	2.2	1.7	57	1.2
02EC009; Holland R at Holland Lndg	2.4	5.3	31	0.0
02HJ001; Jackson Cr at Peterborough	3.3	5.4	69	1.1
02HC028; Little Rouge Cr nr Locust Hill	3.6	6.4	63	1.2
02HC018; Lynde Cr nr Whitby	3.3	5.4	50	1.2
02HD004; NW Ganaraska nr Osaca	2.0	2.2	77	1.3
02HD008; Oshawa Cr at Oshawa	2.3	2.8	40	0.0
02EC018; Pefferlaw Br nr Udora	1.9	2.8	90	1.9
02EC010; Schomberg R nr Schomberg	2.0	6.6	33	0.0
02HC031; W Humber R at Hwy 7	3.1	9.3	33	1.1
02HD009; Wilmot Cr nr Newcastle	2.1	2.3	47	1.7
Mean	2.5	4.3	51	0.9
Standard deviation	0.6	2.1	19	0.7
CV	0.25	0.48	0.37	0.77
Maximum	3.6	9.3	90	2.0
Minimum	1.5	1.7	31	0.0

CV = coefficient of variation.

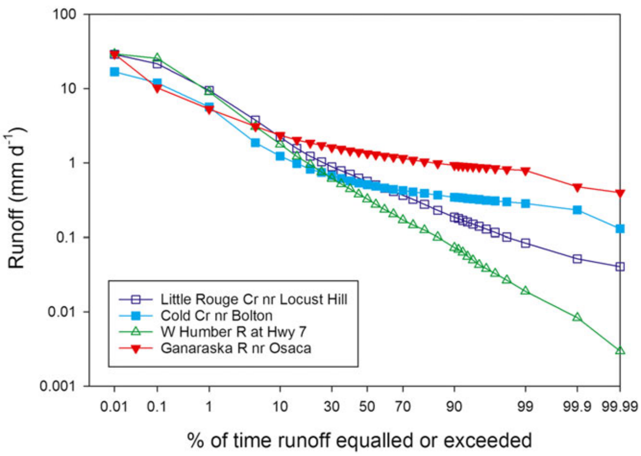

FIGURE 3 Flow duration curves for basins exhibiting the largest and smallest values of Q_5–Q_{95} (02HC028—Little Rouge Cr nr Locust Hill and 02HC023—Cold Creek nr Bolton) and (Q_5–Q_{95})/Q_{50} (02HC031—W Humber R at Hwy 7 and 02HD003—Ganaraska R nr Osaca)

in dynamic storage with increasing recession rate consistent with previous work (e.g. Stoelzle, Stahl, & Weiler, 2013), based on the dependence of dynamic storage estimates on the form of the discharge sensitivity function $g(Q)$ (Kirchner, 2009). Figure 4 shows $-dQ/dt$ versus Q relationships and estimated recession curves for the North West Ganaraska R (dynamic storage = 77 mm) and the Holland R at Holland Landing (dynamic storage = 31 mm) using the $g(Q)$ function obtained for each basin. The Holland R at Holland Landing generally has a greater recession rate and a much smaller range in relative storage, which in turn leads to smaller dynamic storage. This reflects the basin's smaller mean depth to WT, % GFD and % forest cover, and greater % urban cover relative to the North West Ganaraska R (Table 2). Dynamic storage across all basins was uncorrelated with either Q_5–Q_{95} or (Q_5–Q_{95})/Q_{50}.

3.3 | Temporal trends in hydroclimatic drivers

Most MSC stations in the ORM region did not experience significant temporal trends in precipitation-related hydroclimatic metrics, and all mean trends in precipitation-related metrics are not significantly different from 0 (Table 3). Nevertheless, a subset of stations on the southern flank of the ORM showed temporal increases in WY *P* and declines in snowfall: *P* at the WY timescale (Figures 5 and 6) as well as in fall and spring. The most definitive regional hydroclimatic trend was an increase in WY *PET* for five of eight MSC stations (Figure 5). There was no evidence for regional increases in the availability of water for groundwater recharge (*P* – *PET*) at either the summer or WY time scales.

3.4 | Groundwater levels

Most wells showed an increase in water level with time, although two wells (351–1 and 386–1) had either no observable temporal trend or a modest increase over their period of record (Figure 7). The former were not localized but were spread along the length of the ORM (Figure 2). This suggests an increase in recharge along the crest of the ORM during the 2001–2012 period and is consistent with temporal increases in WY *P* indicated by several MSC climate stations noted above.

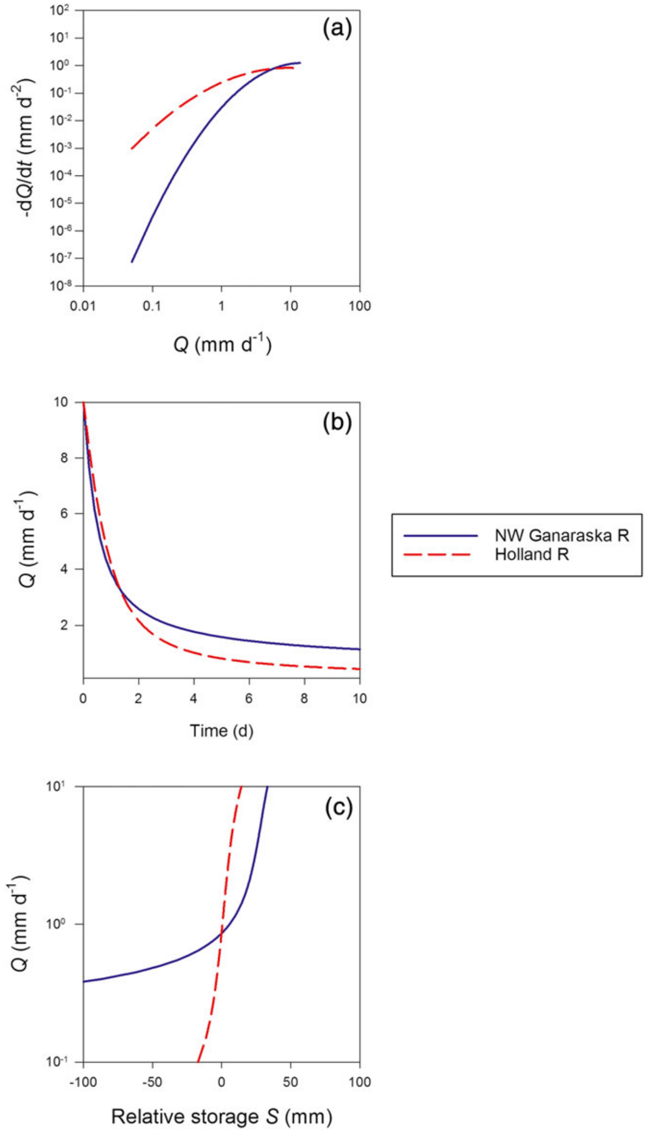

FIGURE 4 (a) $-dQ/dt$ versus Q relationships derived following Kirchner (2009); (b) estimated recession curves; and (c) Q versus relative storage relationships for the North West Ganaraska R and the Holland R at Holland Landing

3.5 | Temporal trends in stream baseflow

Significant ($p < .05$) temporal trends in *BF* for 11 basins gave annual increases of 0.9–2.0 mm year^{-1}, whereas five basins showed no significant trends (Table 2). There was no apparent spatial pattern to *BF* trends, and basins with a significant temporal trend could be found adjacent to others with no trend (Figures 5d and 6b).

3.6 | Principal components analysis

Many basin characteristics exhibited collinearity (e.g., mean basin slope and % GFD), and PCA enabled identification of groups of correlated and normalized basin characteristics for use as independent variables. Three PCs were retained, explaining 87% of the variation in basin characteristics. PC1 explained 48% of total variation and was directly associated with % GFD, mean slope, mean ORMAC thickness, mean depth to WT, and % forest cover and inversely associated with % agricultural cover. PC2 explained 29% of total variation and was directly associated with the coefficient of variation of ORMAC thickness, the coefficient of variation of depth to WT and % water + wetland, and inversely associated with % urban cover. PC3 accounted for 11% of total variation and was directly associated with basin area. Figure 8 presents basin loadings on PCs 1 and 2.

3.7 | Streamflow-based storage metrics versus PCs

Significant inverse relationships were found between both Q_5–Q_{95} and (Q_5–Q_{95})/Q_{50} and PC1 (Figure 9) and to a lesser extent PC2 (data not shown). Slopes of the flow duration curves decreased (i.e., the flow regime became more stable) with increasing thickness of the ORMAC, greater % GFD, increasing basin mean slope, increasing mean depth to WT, greater forest cover, and less agricultural cover. There was also a weaker but nevertheless significant ($p < .05$) relationship between dynamic storage and PC2 (Figure 10). Dynamic storage tended to be greater in basins with more variable ORMAC thickness and depth to WT and more water + wetland coverage.

3.8 | Temporal trends in stream baseflow versus basin characteristics

No significant relationships were found between magnitude of the *BF* temporal trend and the PC scores. There were also no significant relationships between *BF* trend and either Q_5–Q_{95} or (Q_5–Q_{95})/Q_{50}. However, there was a significant ($p < .05$) positive relationship between the magnitude of the *BF* temporal trend and dynamic storage (Figure 11a). The dynamic storage approach assumes that the basin behaves as a first-order dynamical system (Kirchner, 2009), which Birkel, Soulsby, and Tetzlaff (2011) suggested may not apply in basins where a significant amount of precipitation may bypass basin storage as saturation overland flow in valley bottom areas. This implies that the approach may not be appropriate in basins with significant wetland and open water coverage. Most of the study basins had less than 10% water + wetland coverage, whereas this coverage exceeded 15% in four basins (Table 2). Water storage in and release from riparian wetlands and ponds may override the role of groundwater discharge from the ORMAC in controlling streamflow recession and thus the $g(Q)$ function used to derive dynamic storage. Removal of these four basins (Figure 11b) strengthened the positive relationship between magnitude of the *BF* temporal trend and dynamic storage and suggests that basins with larger dynamic storage generally had greater annual change in *BF* over the course of their records.

4 | DISCUSSION

4.1 | Hydroclimatic trends

Temporal changes in precipitation-related hydroclimatic indices for the ORM region were modest. Most MSC stations showing no trends in total *P* or snowfall at the WY or seasonal timescales, and only a subset of stations showed temporal increases in WY *P* and decreases in WY snowfall. These results are broadly consistent with those from previous work. Karl, Groisman, Knight, and Heim Jr (1993) reported minor large-scale increases in annual *P* in southern Canada for the 1950–

TABLE 3 Summary of temporal trends in hydroclimatic drivers in the oak ridges moraine region

Hydroclimatic variable	Total number of stations evaluated	Mean trend ±1 *SD*	Number of stations with positive temporal trends (p < .05)	Range in trend	Number of stations with negative temporal trends (p < .05)	Range in trend
WY precipitation	11	0.7 ± 1.3 mm year^{-1}	3	1.7–3.6 mm year^{-1}	0	-
WY snowfall	9	−0.3 ± 0.5 mm year^{-1}	0	-	3	−0.6 to −1.5 mm year^{-1}
WY snowfall: Precipitation	9	0.000 ± 0.001 year^{-1}	0	-	3	−0.001 to −0.002 year^{-1}
Winter precipitation	11	−0.1 ± 0.3 mm year^{-1}	0	-	1	−0.9 mm year^{-1}
Winter snowfall	9	0.0 ± 0.0 mm year^{-1}	0	-	0	-
Winter snowfall: Precipitation	9	0.000 ± 0.000 year^{-1}	0	-	0	-
Spring precipitation	11	0.1 ± 0.3 mm year^{-1}	1	1.0 mm year^{-1}	0	-
Spring snowfall	9	−0.1 ± 0.1 mm year^{-1}	0	-	2	−0.2 to −0.3 mm year^{-1}
Spring snowfall: Precipitation	9	−0.001 ± 0.001 year^{-1}	0	-	3	−0.002 year^{-1}
Summer precipitation	11	0.0 ± 0.7 mm year^{-1}	2	0.8–1.3 mm year^{-1}	1	−1.6 mm year^{-1}
Fall precipitation	11	0.2 ± 0.4 mm year^{-1}	2	1.1 mm year^{-1}	0	-
Fall snowfall	9	0.0 ± 0.1 mm year^{-1}	0	-	3	−0.1 to −0.2 mm year^{-1}
Fall snowfall: Precipitation	9	0.000 ± 0.000 year^{-1}	0	-	2	−0.001 year^{-1}
Winter precipitation: WY precipitation	11	0.000 ± 0.000 year^{-1}	0	-	2	−0.001 year^{-1}
Spring precipitation: WY precipitation	11	0.000 ± 0.000 year^{-1}	0	-	1	−0.001 year^{-1}
Winter + spring precipitation: WY precipitation	11	0.000 ± 0.001 year^{-1}	0	-	3	−0.001 to −0.002 year^{-1}
WY potential evapotranspiration	8	1.0 ± 0.9 mm year^{-1}	5	1.2–2.4 mm year^{-1}	0	-
WY precipitation—Potential evapotranspiration	8	0	0	-	0	-
Summer precipitation—Potential evapotranspiration	8	0.1 ± 0.2 mm year^{-1}	1	0.7 mm year^{-1}	0	-

SD = standard deviation; WY = water year.

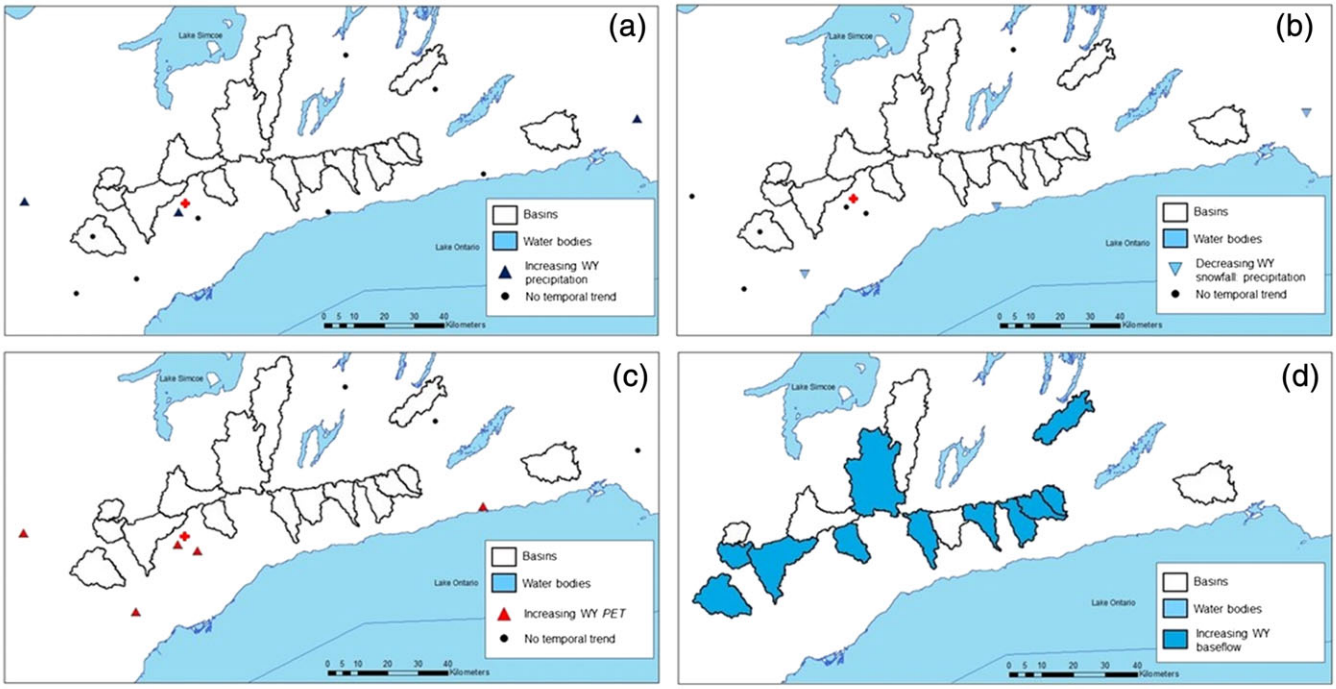

FIGURE 5 (a) Meteorological Service of Canada stations with significant (p < .05) temporal trends in WY P; (b) MSC stations with significant temporal trends in WY snowfall: P; (c) MSC stations with significant temporal trends in WY PET; (d) basins showing increasing temporal trends in WY BF significant at p = .05 (shaded) and no temporal trends (unshaded) for Water Survey of Canada stations with >25 years of record

1990 period, whereas annual rates of change in total P and in snowfall in the Great Lakes—St Lawrence Lowlands region for 1985–1995 (Mekis & Hogg, 1999) were similar in direction and magnitude to those given in Table 3. Vincent et al. (2015) showed significant (p < .05) increases (10–15% over 65 years) in annual P for 1948–2012 for the ORM region. This is equivalent to an increase in annual P of 1.3–2 mm year^{-1} assuming an annual P of 850 mm for the ORM region, similar to values for some MSC stations in Table 3. Mekis and Vincent (2011) reported decreases in annual snowfall for 1959–2009 for stations in the ORM region; however, none were significant at p = .05. The results for the ORM region in Table 3 also indicate either no change or a decrease in the WY snow: P ratio. This corresponds with the annual change in the fraction of solid P to total P of only approximately −0.001 year^{-1} reported by Karl et al. (1993) for 1950–1990 for southern Canada, which was attributed to a slight increase in total P and declines in annual snowfall. There was no conclusive evidence for a change in the seasonality of P in the ORM region, which was supported by Vincent et al. (2015). There was much stronger evidence for increased air temperatures and associated increases in WY PET, which agrees with Kling et al.'s (2003) observation that the historical trends for the Great Lakes region are for greater annual temperatures (and thus increasing PET). It is important to note that although only a subset of MSC stations showed significant temporal increases in WY P, there appears to have been a general increase in groundwater recharge along the crest of the ORM for the 2001–2012 period as indicated by monitoring well records (Figure 7).

4.2 | Streamflow-based storage metrics

Flow duration curve slopes for the ORM basins were indicative of stable flow regimes with relatively little difference between high and low flows and likely reflected the large storage potential of the ORMAC. Data based on flow duration curves from Burn et al. (2008) for basins in Southern Ontario gave an average Q_5–Q_{95} of 6.9 mm day^{-1}, and a mean (Q_5–Q_{95})/Q_{50} of 7.7. Values from the ORM basins tended to be much smaller than these (Table 2), corresponding to role of storage in the ORMAC in mediating the translation of water inputs to streamflow outputs. Dynamic storage estimates for the basins were similar in size to those reported elsewhere (Kirchner, 2009: 107–124 mm; Teuling et al., 2010: 104 mm; Birkel et al., 2011: 40–55 mm).

4.3 | Temporal trends in stream baseflow

Many basins had significant temporal increases in BF, with trends ranging from 0.9 to 2.0 mm year^{-1}. These values compare well with modelled increases in average annual recharge in Grand River basin to the west of the study area in the order of 2.5 mm year^{-1} for 1960–1999 (Jyrkama & Sykes, 2007). They are also consistent with partial evidence of increasing trends in WY P and groundwater recharge along the ORM reported here, and Ficklin et al.'s (2016) observation of good correspondence between temporal trends in BF and P at the national scale in the United States. Any increase in WY P in the ORM region has been partly countered by increasing PET, which is greatest in summer. However, there was no regional temporal trend in the availability of water (P – PET) to recharge basins at the WY or summer time scales to account for the observed increases in BF. Such increases have occurred in the face of regional increases in PET, which might be expected to reduce the availability of P to contribute to groundwater recharge and thus BF. For example, Sultana and Coulibaly (2011) predicted that a 1–10% increase in annual ET for a small basin to the southwest of the ORM would lead to a 0.5–6%

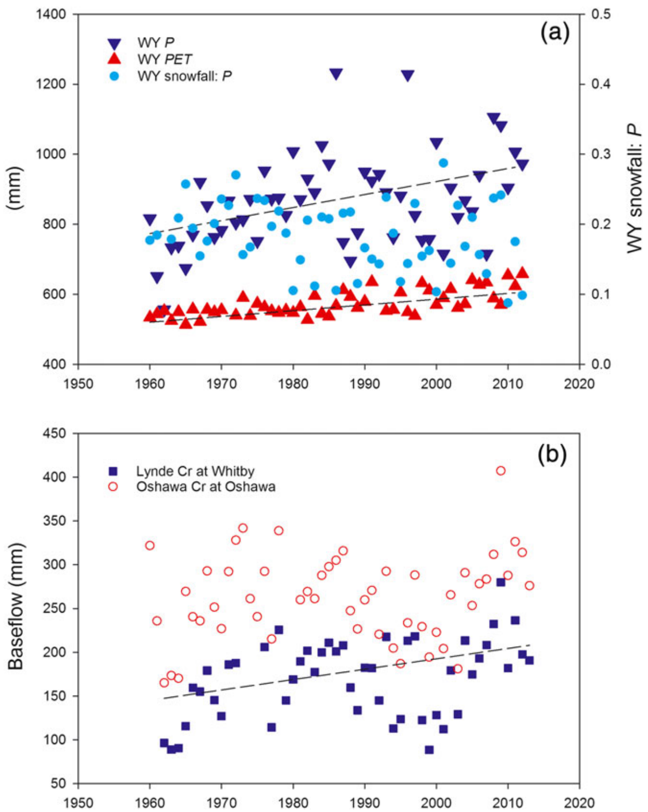

FIGURE 6 (a) Temporal trends and Kendall–Theil best-fit lines for WY *P*, WY *PET*, and WY snowfall: *P* and WY *PET* for Richmond Hill, Ontario (noted with a red cross in Figure 5). Trends for WY *P* and WY *PET* are significant at *p* = .05, whereas the trend for WY snowfall: *P* is not significant. (b) Temporal trends and Kendall–Thiel best-fit lines for WY *BF* for 02HC018—Lynde Cr nr Whitby and the adjacent 02HD008—Oshawa Cr at Oshawa (see Figure 1 and Table 1 for basin locations and characteristics). The trend for Lynde Cr nr Whitby baseflow is significant at *p* = .05, and there is no significant trend for Oshawa Cr at Oshawa baseflow

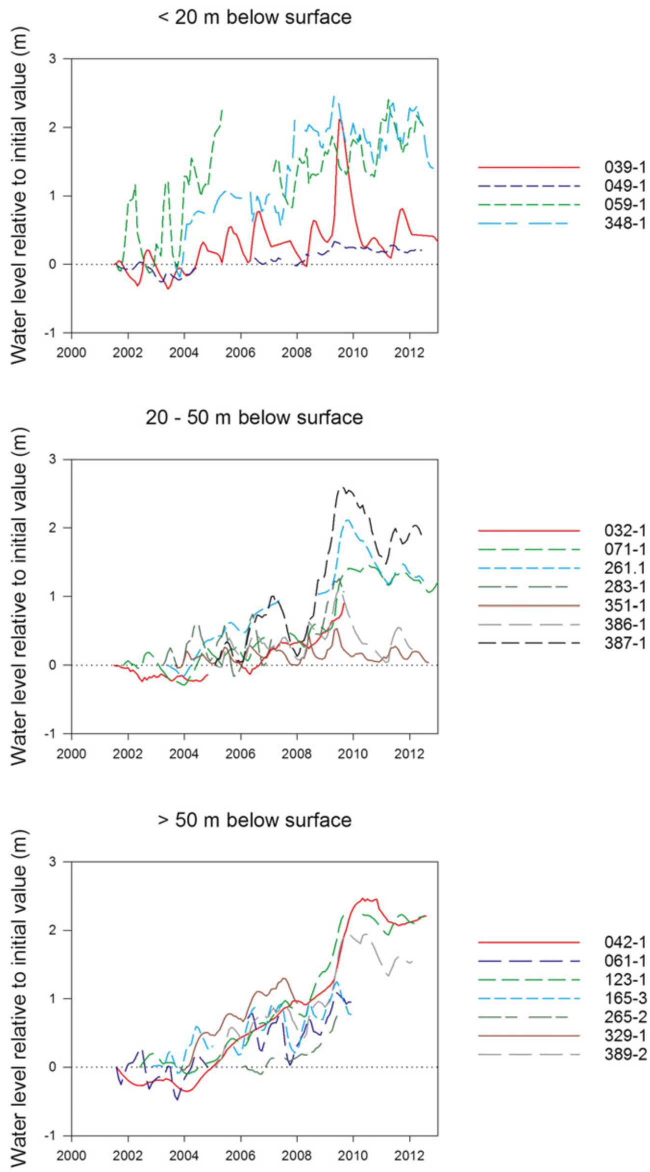

FIGURE 7 Monthly average water levels (expressed relative to the initial monthly average value) for Ontario provincial groundwater monitoring wells located within the ORM planning boundary and screened within the sand and gravel deposits of the ORMAC at different depths below ground surface. See Figure 2 for well locations

decrease in annual groundwater recharge and declines in *BF*. Bialkowski and Buttle (2015) found that 10–50% of *P* during the growing season (May–September) may recharge along the crest of the ORM during relatively wet years. Nevertheless, the main period of recharge on the ORM is during late winter–early spring when *PET* is relatively minimal (Bialkowski & Buttle, 2015), and >60% of mean WY *BF* for the 2006–2012 period for the study basins occurred during the November–April period. Thus, temporal increases in both *BF* and *PET* for the ORM basins were consistent with Shaw, McHardy, and Riha (2013), who found little direct linkage between *ET* and *BF* for basins in the northern United States. Temporal increases in *BF* were likely the result of increased *P* in winter, spring, and/or possibly fall and a greater fraction of this *P* falling as rain rather than snow, as was partly supported by declines in Fall and Spring snowfall and snowfall: *P* for a subset of MSC stations (Table 3). Eckhardt and Ulbrich (2003) suggest that a reduction in the snow: *P* ratio should increase groundwater recharge through increased infiltration in winter. This was supported by simulations of groundwater recharge under warmer winter temperatures in the Grand River basin to the west of the ORM (Jyrkama & Sykes, 2007).

There does not appear to be any association between the locations of MSC stations with significant temporal increases in *P* and basins with significant temporal increases in *BF* (Figure 5), and not all basins showed significant increases in *BF*. This variability in *BF* response occurred across a relatively small region (~14,400 km^2 in size) with similar physiography and geology. Jyrkama and Sykes (2007) noted that simulated groundwater recharge was not uniform across the Grand River basin to the west of the ORM, which they attributed to such factors as groundwater levels, ground surface characteristics, and the nature of subsurface materials. Ficklin et al. (2016) also found instances of basins within regions in the United States that experienced temporal trends in *BF* that differed significantly from the average regional behaviour. This demonstrates that the linkage between predicted changes in hydroclimatic drivers and basin hydrologic response is not a simple one and emphasizes the need to consider basin characteristics when attempting to forecast the hydrologic reaction to climate change.

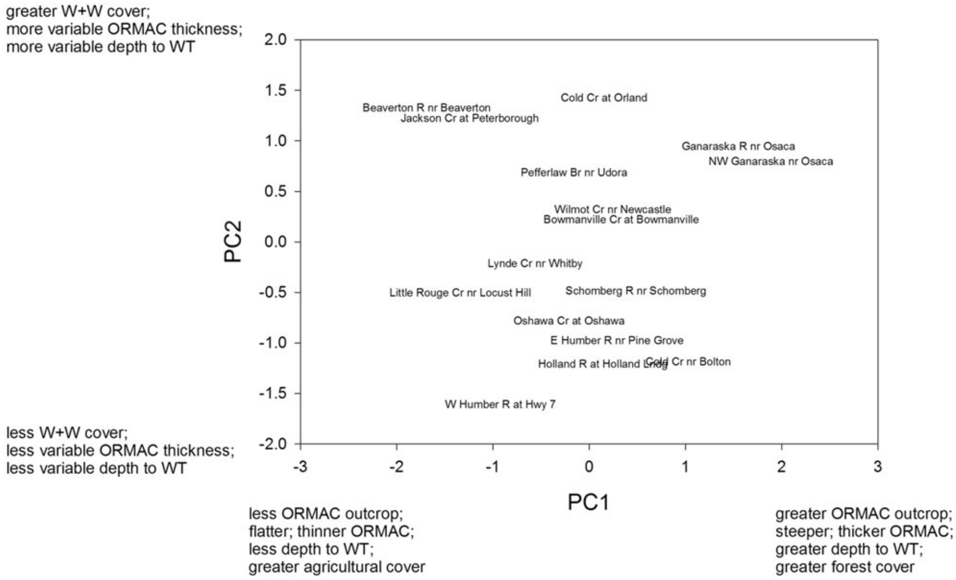

FIGURE 8 Basin loadings on PC1 and PC2 (see Figure 1 and Table 1 for basin locations and characteristics). ORMAC = Oak Ridges Moraine aquifer complex; WT = water table; W + W = water + wetland

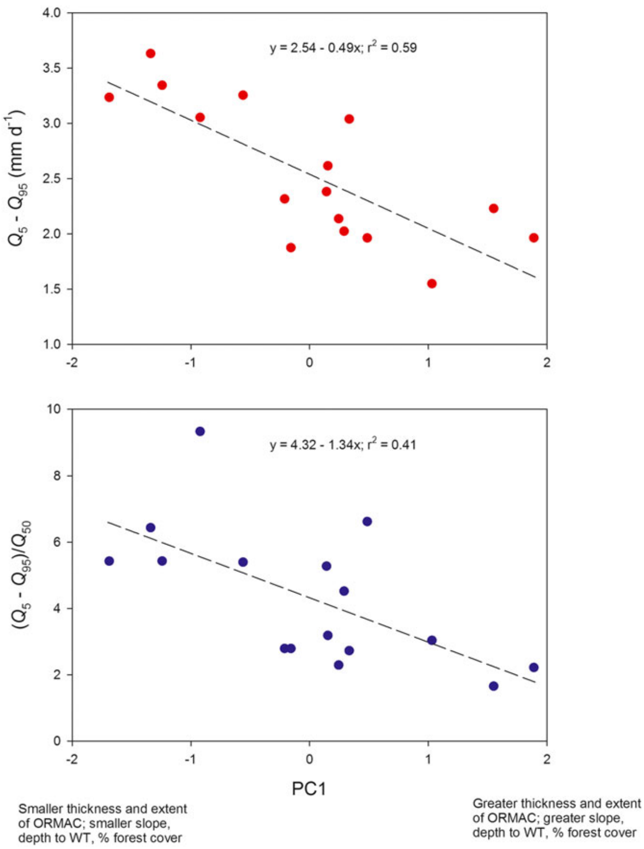

FIGURE 9 Q_5–Q_{95} versus PC1 score (a) and (Q_5–Q_{95})/Q_{50} versus PC1 score (b). Best-fit relationships are significant at p = .05. ORMAC = Oak Ridges Moraine aquifer complex; WT = water table

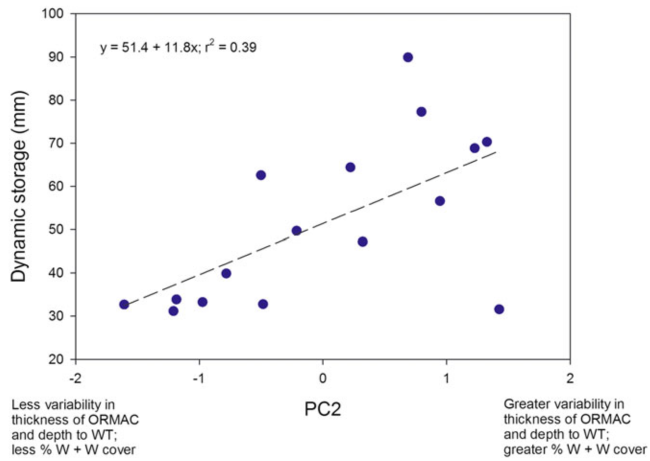

FIGURE 10 Dynamic storage versus PC2 score. Best-fit relationship is significant at p = .05. ORMAC = Oak Ridges Moraine aquifer complex; WT = water table; W + W = water + wetland

4.4 | Basin characteristics, storage metrics, and *BF* temporal trends

There is a long-standing interest in the hydrologic community in explaining inter-basin differences in storage properties by relating the results of recession analyses such as the dynamic storage estimates derived here to a basin's physical characteristics (Stoelzle et al., 2013). Thus, basins draining the ORM with greater thickness and extent of the ORMAC, greater mean basin slope, depth to WT, and forest cover (PC1) were associated with flatter flow duration curves and less variable streamflow regimes (Figure 9). There was also a positive relationship between dynamic storage and basin loading on PC2 (Figure 10). The discharge sensitivity function $g(Q)$ used to estimate dynamic storage (Kirchner, 2009) is based on second-order polynomial best-fit relationships between binned values of $\ln(-dQ/dt)$ and $\ln(Q)$ of the type shown in Figure 4. Relating the parameters of such relationships to basin characteristics using approaches such as PCA (as done here) may allow the dynamic storage approach to be applied to ungauged basins (Adamovic, Braud, Branger, & Kirchner, 2015).

Reasons for the positive association between dynamic storage and PC2 (greater variability in thickness of both the ORMAC and depth to

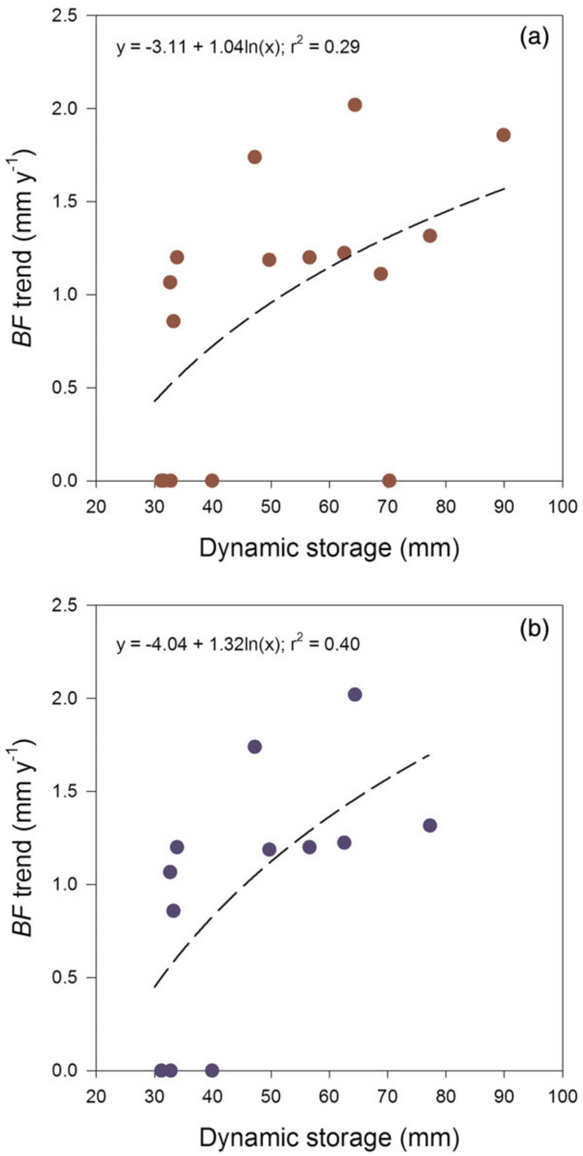

FIGURE 11 Temporal trend in stream baseflow (*BF*) significant at *p* = .05 versus dynamic storage for all basins with >25 years of record (a) and for basins with <10% water + wetland coverage (b). Best-fit relationships are significant at *p* = .05

WT, greater % water + wetland cover) are not obvious at first glance. Basins scoring high on PC2 have areas (e.g., riparian wetlands and zones with water tables close to the ground surface) that may quickly translate precipitation into streamflow, as well as other zones where inputs recharge groundwater within the ORMAC and may be stored for several years before their release to stream channels. Differing rates of water release from these various stores (Stewart, 2015) may lead to a greater range of storage change between high and low streamflows (i.e., greater dynamic storage) relative to basins with more uniform ORMAC thickness and depth to WT and less extensive water + wetland cover.

The relatively weak relationship between dynamic storage and those basin properties represented by PC2 indicates that the basin metrics used here provide an incomplete characterization of the controls on storage and release of water inputs to streamflow. Metrics such as mean aquifer thickness and mean depth to WT provide only a partial picture of the ORMAC's hydrogeology. For example, the Newmarket till aquitard underlying the ORMAC is breached by sediment-filled palaeochannels, which enhance recharge from the ORMAC into deeper confined aquifers (Sharpe et al., 2002). This may result in a portion of groundwater recharge in a given basin leaving as deeper regional groundwater flow and not being captured as *BF* at the WSC station. This issue might be particularly important in smaller headwater basins (Hinton, Russell, Bowen, & Ahad, 1998) and may bias storage–discharge relationships such as the dynamic storage estimates presented here (Buttle, 2016; Krakauer & Temimi, 2011). Conversely, the lower reaches of stream channels in several basins have incised into these confined aquifers, particularly where the WSC station is further down the flank of the ORM (e.g., W Humber R at Hwy 7, Pefferlaw Br nr Udora). Thus, stream *BF* at the WSC stations reflects local groundwater discharge from the ORMAC in headwater reaches and regional groundwater contributions in lower stream reaches (Gerber & Howard, 2002). Future attempts to relate dynamic storage to basin characteristics should incorporate a wider range of topographic and hydrogeologic metrics than those employed here, such as those presented by McGuire et al. (2005) and Gerber and Howard (2002), respectively. In addition, the ORM region is undergoing rapid urban development (Furberg & Ban, 2012) as well as changing agricultural practices such as increased use of tile drainage. The implications of these changes in LULC for streamflow and dynamic storage estimates are unclear and deserve further attention. Nevertheless, these results suggest that inter-basin differences in dynamic storage can be explained, at least in part, by topographic, hydrogeologic, and LULC factors.

4.5 | Basin controls on temporal trends in stream baseflow

Dynamic storage was the only storage metric significantly associated with the magnitude of *BF* trends in the ORM region. There were no relationships between temporal trends in *BF* and either the basin characteristics as represented by the PCs or the form of the basin's flow duration curve. Instead, basins with larger dynamic storage generally had larger temporal trends in *BF*, which in turn implies that they show a greater sensitivity (less resistance, sensu Carey et al., 2010) to temporal changes in hydroclimatic drivers of streamflow in this part of Southern Ontario.

The relationship between dynamic storage and the temporal trend in *BF* reported here appears to be a meaningful one, despite the derivation of both dynamic storage and WY *BF* from the same streamflow time series for the study basins. The $g(Q)$ functions used to estimate dynamic storage were obtained from hydrograph recessions that represented a relatively small fraction (maximum of 9%) of the daily flows for the 2006–2012 period, whereas *BF* was determined using a 1-parameter digital filter hydrograph separation for the entire streamflow record. The initial portions of the hydrograph recessions used to derive the $g(Q)$ functions would have been classed as direct run-off according to the 1-parameter digital filter and would not be considered as part of *BF* as estimated here. Consequently, there was no significant (p = .05) correlation between dynamic storage and either mean WY *BF* or mean WY *BF* index (*BF* relative to total run-off) for the 2006–2012 WY period used to derive dynamic storage. In turn, mean

WY *BF* for the 2006–2012 period was not significantly correlated with the temporal trend in *BF* over periods of 25 years or more for the study basins.

Buttle (2016) showed that dynamic storage for a subset of ORM basins was directly correlated with the ratio of the variability of a stable environmental isotope (δ^2H) in streamflow relative to that in precipitation (i.e., basins with greater dynamic storage showed larger relative temporal variability in streamflow δ^2H). Tetzlaff et al. (2009) reported an inverse relationship between basin mean water transit time (the average amount of time between a water molecule entering and leaving a basin—McGuire & McDonnell, 2006) and the ratio of the standard deviations of a tracer in streamflow relative to that in precipitation. Thus, basins with smaller dynamic storage have longer water transit times, greater capacity to store inputs prior to gradually releasing water to stream channels, and less interannual variability in their streamflow regimes (Buttle, 2016). The direct relationship between dynamic storage and magnitude of the temporal trend in *BF* reported here supports the hypothesis that greater basin storage capacity (as indicated by smaller dynamic storage) reduces the sensitivity of *BF* to interannual changes in hydroclimatic drivers such as *P* and *ET*. It is also consistent with Cooper et al.'s (1995) modelling study that showed that greater aquifer storage reduces *BF* sensitivity to climate change perturbations. The ORMAC appears to have considerable potential to buffer interannual changes in groundwater recharge, and Gerber and Howard (2002) derived water particle transit times of 7 years or more for groundwater discharge from the upper ORMAC to headwater streams in the Duffins Creek basin with headwaters on the ORM. The relationship between dynamic storage and the trend in *BF*, although relatively modest, helps to explain inter-basin differences in *BF* temporal trends across the ORM. Given the potential influence of basin characteristics on dynamic storage values (Figure 10), this suggests that basin properties may help to control a basin's *BF* response to changes in hydroclimatic conditions on the ORM.

5 | CONCLUSIONS

There are indications of modest increases in both precipitation and potential evapotranspiration and decreases in the fraction of total precipitation supplied by snow over the past several decades in the ORM region. These changes, along with more recent evidence in increased groundwater recharge along the crest of the ORM, appear to have increased the amount of *BF* discharged from basins draining the ORM. However, these increases in stream *BF* have been non-uniform, with some basins showing marked increases over their period of record, whereas others exhibiting no significant temporal trend. There has been some success in relating metrics of water storage in the ORM basins to basin physiographic, hydrogeologic, and LULC characteristics, such that it may be possible to predict these storage metrics for ungauged basins in the region. However, the absence of a significant correlation between the slope of a basin's flow duration curve and its corresponding dynamic storage estimate implies that these metrics capture different aspects of water storage in and release from the ORM basins, and this issue deserves further study. Nevertheless, results suggest that basin storage may play a role in mediating *BF* reaction to temporal changes in hydroclimatic drivers of streamflow and may partly account for inter-basin differences in that response. This in turn reinforces the need to consider basin characteristics such as those related to water storage and release in attempts to forecast regional streamflow responses to climate change.

ACKNOWLEDGMENTS

This work was supported by the Natural Sciences and Engineering Research Council of Canada. Thanks to Kayla Goguen, Wes Greenwood, and Ciara Cooke for data analysis; Rick Gerber for hydrogeologic data; Cathy Eimers and two anonymous reviewers for constructive comments on an earlier version of this paper; and James Kirchner for providing the initial impetus for this work.

ORCID

James M. Buttle http://orcid.org/0000-0001-7231-8972

REFERENCES

Adamovic, M., Braud, I., Branger, F., & Kirchner, J. W. (2015). Assessing the simple dynamical systems approach in a Mediterranean context: Application to the Ardèche catchment (France). *Hydrology and Earth. Systems Science*, *19*, 2427–2449.

Barnett, T. P., Adam, J. C., & Lettenmaier, D. P. (2005). Potential impacts of a warming climate onwater availability in snow-dominated regions. *Nature*, *438*, 303–309.

Berghuijs, W. R., Woods, R. A., & Hrachowitz, M. (2014). A precipitation shift from snow towards rain leads to a decrease in streamflow. *Nature Climate Change*, *4*, 583–586.

Bialkowski, R., & Buttle, J. M. (2015). Soil water recharge for grassed and forested land covers on the oak ridges moraine, Southern Ontario, Canada. *Hydrology Research*, *47*, 390–408.

Birkel, C., Soulsby, C., & Tetzlaff, D. (2011). Modelling catchment-scale water storage dynamics: Reconciling dynamic storage with tracer-inferred passive storage. *Hydrological Processes*, *25*, 3924–3936.

Burn, D. H., Buttle, J. M., Caissie, D., MacCulloch, G., Spence, C., & Stahl, K. (2008). The processes, patterns and impacts of low flows across Canada. *Canadian Water Resources Journal*, *33*, 107–124.

Buttle, J. M. (2011). Streamflow response to headwater reforestation in the Ganaraska River basin, Southern Ontario, Canada. *Hydrological Processes*, *25*, 3030–3041.

Buttle, J. M. (2016). Dynamic storage: A potential metric of inter-basin differences in storage properties. *Hydrological Processes*, *30*, 4644–4653.

Buttle, J. M., Greenwood, W. J., & Gerber, R. E. (2015). Spatiotemporal patterns of baseflow metrics for basins draining the oak ridges moraine, Southern Ontario, Canada. *Canadian Water Resources Journal*, *40*, 3–22.

Carey, S. K., Tetzlaff, D., Seibert, J., Soulsby, C., Buttle, J., Laudon, H., … Pomeroy, J. W. (2010). Inter-comparison of hydro-climatic regimes across northern catchments: Synchronicity, resistance and resilience. *Hydrological Processes*, *24*, 3591–3602.

Cheng, C. S., Li, Q., Li, G., & Auld, H. (2012). Possible impacts of climate change on daily streamflow and extremes at local scale in Ontario, Canada. Part II: Future projection. *Atmospheric and Climate Sciences*, *2*, 427–440.

Cherkauer, K. A., & Sinha, T. (2010). Hydrologic impacts of projected future climate change in the Lake Michigan region. *Journal of Great Lakes Research*, *36*, 33–50.

Clarke, C., Hulley, M., Marsalek, J., & Watt, E. (2011). Stationarity of AMAX series of short-duration rainfall for long-term Canadian stations: Detection of jumps and trends. *Canadian Journal of Civil Engineering*, *38*, 1175–1184.

Cooper, D. M., Wilkinson, W. B., & Arnell, N. W. (1995). The effects of climate changes on aquifer storage and river baseflow. *Hydrological Sciences Journal*, *40*, 615–631.

Delidjakova, K., Bello, R., & MacMillan, G. (2014). Measurement of evapotranspiration across different land cover types in the greater Toronto area. Sustainable Technologies Evaluation Program, Toronto and Region Conservation Authority, Toronto, Ontario. 31 p.

Dingman, S. L. (2002). *Physical hydrology* (2nd ed.). (pp. 646). Upper Saddle River, NJ: Prentice-Hall p.

Eckhardt, K., & Ulbrich, U. (2003). Potential impacts of climate change on groundwater recharge and streamflow in a central European low mountain range. *Journal of Hydrology*, *284*, 244–252.

Ficklin, D. L., Robeson, S. M., & Knouft, J. H. (2016). Impacts of recent climate change on trends in baseflow and stormflow in United States watersheds. *Geophysical Research Letters*, *43*, 5079–5088.

Furberg, D., & Ban, Y. (2012). Satellite monitoring of urban sprawl and assessment of its potential environmental impact in the greater Toronto area between 1985 and 2005. *Environmental Management*, *50*, 1068–1088.

Gerber, R. E., & Howard, K. (2000). Recharge through a regional till aquitard: Three-dimensional flow model water balance approach. *Ground Water*, *38*, 410–422.

Gerber, R. E., & Howard, K. (2002). Hydrogeology of the oak ridges moraine aquifer system: Implications for protection and management from the Duffins Creek watershed. *Canadian Journal of Earth Sciences*, *39*, 1333–1348.

Granato, G. E. (2006). Kendall-Thiell Robust Line (KTRLine-version 1.0): A visual basic program for calculating and graphing robust nonparametric estimates of linear relations between two continuous variables. In *U.S. Geological Survey Techniques and Methods Report, Book 4—Hydrological analysis and interpretation, Section A—Statistical analysis, Chapter 7* (pp. 31). Reston, VA: U.S. Geological Survey p.

Greenwood, W. H., & Buttle, J. M. (2014). Effects of reforestation on near-surface saturated hydraulic conductivity in a managed forest landscape, Southern Ontario, Canada. *Ecohydrology*, *7*, 45–55.

Hayhoe, K., Wake, C. P., Huntington, T. G., Luo, L., Schwartz, M. D., Sheffield, J., … Wolfe, D. (2007). Past and future changes in climate and hydrological indicators in the US northeast. *Climate Dynamics*, *28*, 381–407.

Hinton, M. J., Russell, H. A. J., Bowen, G. S., & Ahad, J. M. E. (1998). Groundwater discharge in the Humber River watershed. In A. R. Piggott (Ed.), *Groundwater in a watershed context: Symposium proceedings* (pp. 213–220). Cambridge, ON: Canadian Water Resources Association.

Holysh, S., & Gerber, R. (2014). Groundwater knowledge management for Southern Ontario: An example from the oak ridges moraine. *Canadian Water Resources Journal*, *39*, 240–253.

Hubbart, J. A., & Zell, C. (2013). Considering streamflow trend analyses uncertainty in urbanizing watersheds: A baseflow case study in the central United States. *Earth Interactions*, *17*, 1–28.

Jyrkama, M. I., & Sykes, J. F. (2007). The impact of climate change on spatially varying groundwater recharge in the Grand River watershed Ontario. *Journal of Hydrology*, *338*, 237–250.

Karl, T. R., Groisman, P. Y., Knight, R. W., & Heim, R. R. Jr. (1993). Recent variations of snow cover and snowfall in North America and their relation to precipitation and temperature variations. *Journal of Climate*, *6*, 1327–1344.

Kirchner, J. W. (2009). Catchments as simple dynamical systems: Catchment characterization, rainfall-runoff modeling, and doing hydrology backward. *Water Resources Research*, *45*, W02429. https://doi.org/10.1029/2008WR006912.

Kling, G. W., Hayhoe, K., Johnson, L. B., Magnuson, J. J., Polasky, S., Robinson, S. K., … Wilson, M. L. (2003). *Confronting climate change in the Great Lakes region: Impacts on our communities and ecosystems*. (pp. 92). Washington, D.C: Union of Concerned Scientists, Cambridge, Massachusetts, and ecological Society of America p.

Krakauer, N. Y., & Temimi, M. (2011). Stream recession curves and storage variability in small watershed. *Hydrologic and Earth System Sciences*, *15*, 2377–2389.

Lim, K. H., Engel, B. C., Tang, Z., Choi, J., Kim, K.-S., Muthurkrishnan, S., & Tripathy, D. (2005). Automated web GIS based hydrograph analysis tool, WHAT. *Journal of the American Water Resources Association*, *41*, 1407–1416.

McGuire, K. J., & McDonnell, J. J. (2006). A review and evaluation of catchment transit time modeling. *Journal of Hydrology*, *350*, 543–563.

McGuire, K. J., McDonnell, J. J., Weiler, M., Kendall, C., McGlynn, B. L., Welker, J. M., & Seibert, J. (2005). The role of topography on catchment-scale water residence time. *Water Resources Research*, *41*, W05002. https://doi.org/10.1029/2004WR003657.

McLean, R. K., & Watt, W. E. (2005). Regional low flow frequency relations for Central Ontario. *Canadian Water Resources Journal*, *30*, 179–196.

McNamara, J. P., Tetzlaff, D., Bishop, K., Soulsby, C., Seyfried, M., Peters, N. E., … Hooper, R. (2011). Storage as a metric of catchment comparison. *Hydrological Processes*, *25*, 3364–3371.

Mekis, E., & Hogg, W. D. (1999). Rehabilitation and analysis of Canadian daily precipitation time series. *Atmosphere-Ocean*, *37*(1), 53–85.

Mekis, E., & Vincent, L. A. (2011). An overview of the second generation adjusted daily precipitation dataset for trend analysis in Canada. *Atmosphere-Ocean*, *49*, 163–177.

Meyer, S. C. (2005). Analysis of base flow trends in urban streams, Northeastern Illinois, USA. *Hydrogeology Journal*, *13*, 871–885.

Moin, S. M. A., & Shaw, M. A. (1985). *Regional flood frequency analysis for Ontario streams. Vol. 1. Single station analysis and index method*. Environment Canada, Burlington, ON: Inland Waters Directorate.

Ontario Geological Survey (OGS). (2003). Surficial Geology of Southern Ontario. Project Summary and Technical Document. Miscellaneous Release Data: 128. Ministry of Northern Development and Mines, Sudbury, ON.

Ontario Ministry of Natural Resources (OMNR). (2006). Digital elevation models—Version 2.0.0- Provincial Tiled Dataset. Ontario Ministry of Natural Resources, Peterborough, ON [raster file].

Price, K. (2011). Effects of watershed topography, soils, land use, and climate on baseflow hydrology in humid regions: A review. *Progress in Physical Geography*, *35*, 465–492.

Rivard, C., Vigneault, H., Piggott, A. R., Larocque, M., & Anctil, F. (2009). Groundwater recharge trends in Canada. *Canadian Journal of Earth Sciences*, *46*, 841–854.

Sawicz, K., Wagener, T., Sivapalan, M., Troch, P. A., & Carrillo, G. (2011). Catchment classification: Empirical analysis of hydrologic similarity based on catchment function in the eastern USA. *Hydrology and Earth Systems Science*, *15*, 2895–2911.

Sharpe, D. R., Hinton, M. J., Russell, H. A. P., & Desbarats, A. J. (2002). The need for basin analysis in regional hydrogeological studies: Oak ridges moraine, Southern Ontario. *Geoscience Canada*, *29*, 3–20.

Sharpe, D. R., Russell, H. A. J., & Logan, C. (2007). A 3-dimensional geological model of the oak ridges moraine area, Ontario, Canada. *Journal of Maps*, *3*, 239–253.

Shaw, S. B., McHardy, T. M., & Riha, S. J. (2013). Evaluating the influence of watershed moisture storage on variations in base flow recession rates during prolonged rain-free periods in medium-sized catchments in New York and Illinois, USA. *Water Resources Research*, *49*, 6022–6028.

Singh, R., Wagener, T., Werkhoven, K. V., Mann, M. E., & Crane, R. (2011). A trading-space-for-time approach to probabilistic continuous streamflow predictions in a changing climate-accounting for changing watershed behavior. *Hydrology and Earth Systems Science Discussions*, *15*, 3591–3603.

SOLRIS Technical team. (2008). Southern Ontario Land Resource Information System (SOLRIS), Phase 2—Data Specifications Version 1.2. Ontario Ministry of Natural Resources internal publication, Peterborough, ON, 17 p.

Soulsby, C., Piegat, K., Seibert, J., & Tetzlaff (2011). Catchment-scale estimate of flow path partitioning and water storage basin on transit time and runoff modelling. *Hydrological Processes*, *25*, 3960–3976.

Stewart, M. K. (2015). Promising new baseflow separation and recession analysis methods applied to streamflow at Glendhu catchment, New Zealand. *Hydrology and Earth Systems Science*, *19*, 2587–2603.

Stoelzle, M., Stahl, K., & Weiler, M. (2013). Are streamflow recession characteristics really characteristic? *Hydrology and Earth Systems Science*, *17*, 817–828.

Stuckey, M. H. (2006). Low-flow, base-flow, and mean-flow regression equations for Pennsylvania streams: U.S. Geological Survey Scientific Investigations Report 2006–5130, 84 p.

Sultana, Z., & Coulibaly, P. (2011). Distributed modelling of future changes in hydrological processes of Spencer Creek watershed. *Hydrological Processes*, *25*, 1254–1270.

Tetzlaff, D., Seibert, J., McGuire, K. J., Laudon, H., Burns, D. A., Dunn, S. M., & Soulsby, C. (2009). How does landscape structure influence catchment transit time across different geomorphic provinces? *Hydrological Processes*, *23*, 945–953.

Teuling, A. J., Lehner, I., Kirchner, J. W., & Seneviratne, S. I. (2010). Catchments as simple dynamical systems: Experience from a Swiss prealpine catchment. *Water Resources Research*, *46*, W10502. https://doi.org/10.1029/2009WR008777.

Vincent, L. A., Zhang, X., Brown, R. D., Feng, Y., Mekis, E., Milewska, E. J., … Wang, X. L. (2015). Observed trends in Canada's climate and influence of low-frequency variability modes. *Journal of Climate*, *28*, 4545–4560.

How to cite this article: Buttle JM. Mediating stream baseflow response to climate change: The role of basin storage. *Hydrological Processes*. 2018;32:363–378. https://doi.org/10.1002/hyp.11418