14 Hydrological Processes of Reference Watersheds in Experimental Forests, USA

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14.1 Introduction

Long-term research at small, gauged, forested watersheds within the USDA Forest Service, Experimental Forest and Range network (USDA-EFR) has contributed substantially to our current understanding of relationships between forests and streamflow (Vose *et al.*, 2014). Many of these watershed studies were established in the early to mid-20th century and have been used to evaluate the effects of forest disturbances such as harvesting, road construction, wild and prescribed fire, invasive species and changes in tree species composition on hydrological responses including stormflows, peak flows, water yield, ground water table and evapotranspiration. Forest hydrologists and natural resources managers are still working to fully understand the effects of watershed disturbances on hydrology, water quality and other ecosystem services (Zegre, 2008). Much of our knowledge on this topic is derived from steep, mountainous watersheds where these studies were initially conducted. An assessment by Sun *et al.* (2002) has shown that low-gradient watersheds with forested wetlands generally have lower water yields, lower runoff ratios and higher evapotranspiration than upland-dominated watersheds, adding to our knowledge of forest hydrology, particularly on the effects of topography on streamflow patterns and stormflow peaks and volumes.

While paired watershed studies (Bosch and Hewlett, 1982; Brown *et al.*, 2005) have been invaluable in understanding the hydrological response to disturbances, reference watersheds can provide valuable insight into hydrological processes in relatively undisturbed forest ecosystems. The term 'reference' watershed is favoured over the term 'control' because reference watersheds also change over time in response to natural (e.g. windthrow, insects, fire, hurricanes, climatic extremes) and human-induced disturbances (e.g. atmospheric pollution, invasive species, climate change). However, reference watersheds experience disturbances that are typically minor compared with most experimental treatments. Several recent studies have synthesized data from small reference watersheds, including

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those in the USDA-EFR network, highlighting important insights that can be gained from long-term data (Jones *et al*., 2012; Argerich *et al*., 2013; Creed *et al*., 2014).

This chapter provides an overview and comparison of factors influencing hydrological processes, especially streamflow dynamics and evapotranspiration, at ten relatively undisturbed reference watersheds in the USDA-EFR network (Fig. 14.1, Table 14.1). We demonstrate the breadth of the hydrogeological, topographic, climatic and ecological characteristics of reference watersheds by discussing how factors such as climate, topography, geology, soils and vegetation influence runoff generation (Fig. 9.1, Chapter 9, this volume) of these reference watersheds. We also briefly consider how site factors influence evapotranspiration, which determines water balance and regulates streamflow. This enhances our current understanding of the hydrological behaviour of these watersheds enabling us to better predict responses to, and prepare for, future management and environmental changes (Jones *et al*., 2009; Vose *et al*., 2014).

Located in vastly different ecohydrological regions, these watersheds have multiple factors influencing the streamflow (*Q*) regimes. Therefore we chose to assess differences in streamflow magnitudes and frequencies using flow duration curves (FDCs) and their flow percentiles (Searcy, 1959; Vogel and Fennessey, 1995). FDCs have been used to study integrated streamflow responses to different types and distributions of storm runoff events (i.e. rainstorms, snowmelt) and landscape characteristics, and have been applied extensively to evaluate streamflow responses to changing climate and other disturbances (e.g. Arora and Boer, 2001). An FDC with a steep slope throughout indicates a stream with more variable flow, whereas a flat slope is indicative of more stable flow with less variability. A steep slope at the upper end indicates more flashy streams with direct runoff characterizing a flood regime, while a flatter slope indicates

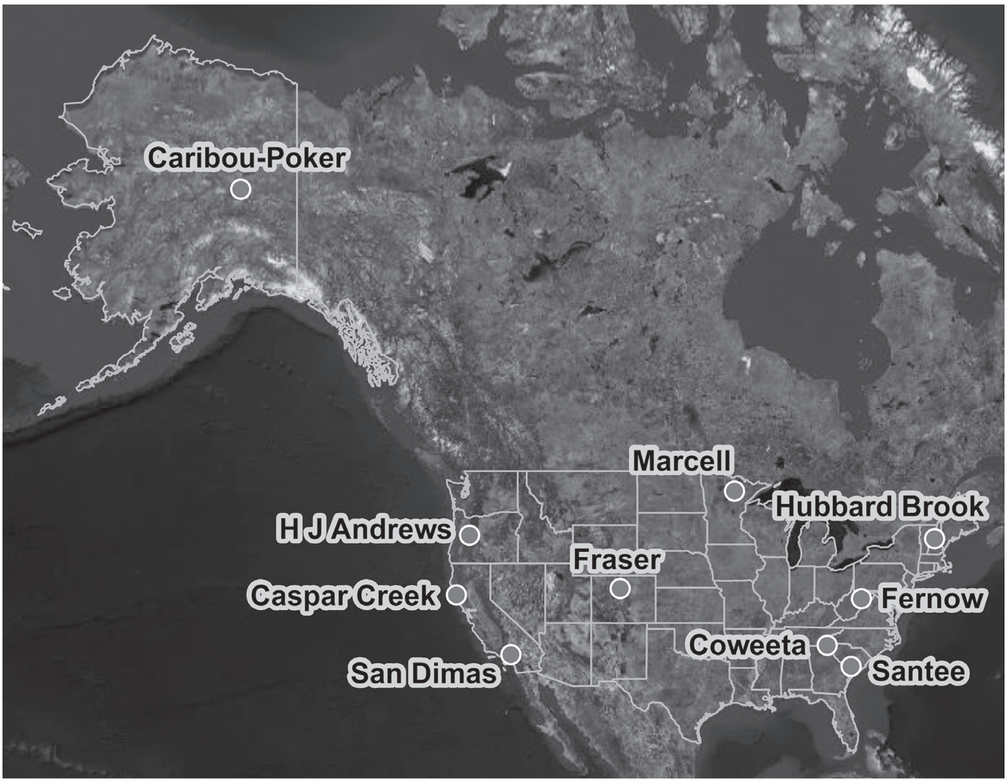

Fig. 14.1. Map of the ten USDA Forest Service Experimental Forests included in this chapter.

Table 14.1. Comparative characteristics of reference watersheds at ten long-term paired experimental forest watersheds in the USA.

Watershed characteristics	Caribou-Poker (CPCRW), Alaska	Caspar Creek (CCEW), California	Coweeta (CHL), North Carolina	Fernow (FnEF), West Virginia	Fraser (FrEF), Colorado	H.J. Andrews (HJAEF), Oregon	Hubbard Brook (HBEF), New Hampshire	Marcell (MEF), Minnesota	San Dimas (SDEF), California	Santee (SEF), South Carolina
Physiographical region as per classification by Fenneman	Yukon–Tanana Northern Plateaus Province (12)	Pacific Mountain System, 23f, Pacific Border Province	Appalachian Highlands, 5b, Blue Ridge Province	Appalachian Highlands, 8c, Appalachian Plateau	Rocky Mountain Systems, 15, Southern Rocky Mountain	Pacific Mountain System, 22b, Sierra-Cascade Mountain	Appalachian Highlands, 9b, New England Province	12b, Interior Plain, Central Lowland	Pacific Mountain System, 23g, Pacific Border Province	Atlantic Plain Coastal Plain
Climatic region as per classification by Köppen (Peel *et al.*, 2007)	Dfc, continental subarctic or boreal taiga	Csb, temperate/ mesothermal, Mediterranean	Cfb, marine temperate	Dfb, continental warm summer	Dsc, continental cold winter & cool, short, dry summer	Csb, temperate/ mesothermal dry summer	Dfb, continental warm summer	Dfb, continental warm summer	Csa, Mediterra-nean hot summer	Cfa, temperate humid subtropical
Watershed #/ year gauging started	(CPCRW – C2) 1969	(North Fork, NF) 1962	(WS18) 1936	(WS4) 1951	(East St Louis, ESL) 1943	(WS02) 1952	(WS 3) 1957	(S2) 1961	(Bell 3) 1938	(WS80) 1968
Latitude/ longitude	65.17°N, 147.50°W	39.35°N, 123.73°W	35.05°N, 83.43° W	39.03°N, 79.67°W	39.89°N, 105.88°W	44.21°N, 122.23°W,	44.0°N, 71.7°W	47.514°N, 93.473°W	34.20°N, 117.78°W	33.17°N, 79.77°W
Elevation (m amsl)[a]	210–826	30–322	726–993	670–866	2907–3719	572–1079	527–732	420–430	755–1080	3.7–10
Average slope (%)	31	49	52	20	16	41	21	3	34	<3
Drainage area (ha)	520	473	12.5	38.7	803	61	42.4	9.7	25	160
Vegetation type/ average leaf area index (LAI) (m^2/m^2)	Boreal forest/ LAI = 4.1	Second-growth coast redwood/ Douglas fir forest/ LAI = 11.7	Mixed deciduous forest/ LAI = 6.2	Mixed deciduous hardwoods/ LAI = 4.5	Mixed spruce/fir and pine/ LAI = 3.44	Conifer primarily Douglas fir and western hemlock/ LAI = 12	Northern hardwood/ LAI = 6.3	Deciduous uplands; black spruce–*spaghnum* bog/ LAI = N/A	Mixed chaparral/ LAI = 2.2	Pine mixed hardwood/ LAI = 2.8
Dominant geology/ aquifer	Yukon–Tanana metamorphic complex/ discontinuous permafrost	Marine shales & sandstones, Coastal Belt of the Franciscan Complex	Quartz dioritic gneiss predominant, Coweeta Group	Sedimentary; Hampshire formation sandstone and shales	Gneiss and schist, glacial till	Volcanic tuffs and breccias covered with andesite colluvium	Meta-sedimentary/ mica schist, calc-silicate granulite, Silurian Rangeley formation	Glacial till overlying deep outwash sands above Precambrian bedrock	Precambrian metamor-phics and Mesozoic granitics	Sedimentary/ Santee limestone

Continued

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Watershed characteristics	Caribou-Poker (CPCRW), Alaska	Caspar Creek (CCEW), California	Coweeta (CHL), North Carolina	Fernow (FnEF), West Virginia	Fraser (FrEF), Colorado	H.J. Andrews (HJAEF), Oregon	Hubbard Brook (HBEF), New Hampshire	Marcell (MEF), Minnesota	San Dimas (SDEF), California	Santee (SEF), South Carolina
Dominant soil type/ depth	Olnes Silt Loam – *Typic Cryorthents*; Fairplay Silt Loam – *Fluvaquentic Endoaquolls*; Ester Silt Loam – *Typic Histoturbels*/ 0.2–0.5 m	Vandamme Series Ultisols (*Typic Haplohumults*)/ 1–1.5 m	Coweeta–Evard–Saunook complex (fine-loamy, mixed, *Mesic* and *Humic Hapludults*)/ >1.5 m	Loamy-skeletal, mixed mesic *Typic Dystrudepts*/ 1 m	Leighcan Series, loamy-skeletal, *Typic Dystrocyepts*/ <1.5 m	50% andesite colluvium, (unnamed soil series), 20% Limberlost series, loam to green breccia/ up to 1.2 m	Lyman–Tun-bridge–Becket Series, Typic *Haplorthods*/ C horizon depth up to 9 m	Warba Series fine loamy, mixed, superactive, frigid haplic Glossudalfs (0.5 m); Loxley peat *Dysic*, frigid *Typic Haplosaprists* (≤7 m)	Trigo–Exchequer Series loamy, mixed, thermic, shallow, *Typic Xerorthents*/ 0.1–0.5 m	Wahee Series clayey, mixed, thermic *Aeric Och-raquults*/ 1.5 m
Long-term mean precipitation (mm)	412	1316	2010[b]	1458	750	2300	1350	780	715	1370
Long-term mean temperature (°C)	–3.0	10.7	12.9[b]	9.3	1.0	8.4	5.9	3.4	14.4	18.3
Long-term mean potential evapotrans-piration (PET) (mm)	466 (Thornth-waite)[c]	660 (Thornth-waite)[c]	1,013 (Hamon)[d]	560 (pan)[e]	383 (Thornth-waite)[c]	546 (Thornth-waite)[c]	550 (Thornth-waite)[c]	552 (Hamon)[d]	753 (Thornth-waite)[c]	967 (Thornth-waite)[c]
Dryness index (DI)	1.13	0.50	0.50	0.38	0.51	0.24	0.41	0.71	1.05	0.71
Long-term mean streamflow (mm)	80 (1978–2003)	659	997	640 (1951–1990)[b]	337	1321	860	170	84	280
Period of streamflow record	1969–present	1962–present	1936–present	1951–present	1943–present	Nov 1952–present	1957–present	1961–present	1938–1960; 1964–2001; 2013–present	1969–1981; 1989–1999; 2003–present

Surface runoff/ flow generation	Saturation excess flow	Infiltration-excess overland flow limited to compacted surfaces	Rare surface runoff, direct channel and fast shallow subsurface flow from VSA[f]	Minimal surface runoff	Rare, only during snowmelt	Minimal surface runoff – high porosity	Minimal surface runoff	Infiltration excess over frozen & saturation excess flow over unfrozen soils	Rare hillslope flow except after fire when infiltration excess flow	Saturation excess flow
Subsurface flow/ drainage	Shallow subsurface flow	Transient subsurface stormflow and soil pipe preferential flow	Shallow lateral flow via soils with high conductivity	Lateral subsurface flow to channel	Shallow subsurface (macropores, coarse soils) and ground-water	Shallow subsurface lateral flow	Lateral subsurface flow	Shallow subsurface with some seepage to an underlying groundwater aquifer	Groundwater flow unknown but presumably high rate of shallow lateral flow	Shallow lateral subsurface flow with negligible deep seepage
Average water table dynamics/ depth (m)	Unknown	1–8 m	>1.5 m except near stream	Unknown	Unknown	Unknown	Variable water table depth	~0.3 m in the bog; 0.5 m in uplands	Unknown; potentially very deep	Shallow, 0.9 m
Riparian areas for hydrology	None	1%	Limited due to steep topography	Limited due to steep topography	Limited to valleys, fens, bogs	Limited	Limited due to steep topography	33% of area is a peatland	~2 %	15% area
Major or extreme natural disturbance	1967 Fairbanks Flood	Windstorm (1995); landslide (2006)	Chestnut blight (1920s–1930s); drought; hurricanes; hemlock wooly adelgid (2003–present)	Chestnut blight; hurricanes; windstorms; SuperStorm Sandy (2012)	Pine bark beetle epidemic	None in reference	Hurricane (1938); ice storm (1998)	Peatland wildfire (1864); potential for derecho, tornados, wildfires	Wildfire	Hurricane Hugo (1989)
Other specific hydrological features	3% permafrost underlain	Fog input, soil pipes			Snowmelt-dominated hydrological regime		Discontinuous dense pan C horizon	Drainage from bog dome and uplands; some deep seepage to the aquifer	Extremely high levels of nitrate from chronic air pollution	Compared with pre-Hugo, flow reversal in paired watersheds after Hugo
Key publica-tion(s) on forest hydrological processes	Haugen *et al*. (1982); Hinzman *et al*. (2002)	Ziemer (1998); Reid and Lewis (2009)	Hewlett and Hibbert (1967); Swift *et al*. (1988)	Reinhart *et al*. (1963); Adams *et al*. (1994)	Alexander *et al*. (1985); Troendle and King (1985)	Rothacher *et al*. (1967); Post and Jones (2001)	Detty and McGuire (2010b); Gannon *et al*. (2014)	Sebestyen *et al*. (2011); Verry *et al*. (2011)	Riggan *et al*. (1985); Meixner and Wohlgemuth (2003)	Harder *et al*. (2007); Jayakaran *et al*. (2014)

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Watershed characteristics	Caribou-Poker (CPCRW), Alaska	Caspar Creek (CCEW), California	Coweeta (CHL), North Carolina	Fernow (FnEF), West Virginia	Fraser (FrEF), Colorado	H.J. Andrews (HJAEF), Oregon	Hubbard Brook (HBEF), New Hampshire	Marcell (MEF), Minnesota	San Dimas (SDEF), California	Santee (SEF), South Carolina
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Experimental forest website	http://www.lter.uaf.edu/bnz_cpcrw.cfm	http://www.fs.fed.us/psw/topics/water/caspar	http://www.srs.fs.usda.gov/coweeta/	http://www.nrs.fs.fed.us/ef/locations/wv/fernow	http://www.fs.usda.gov/efr/fraser	http://andrewsforest.oregonstate.edu	http://www.hubbardbrook.org	http://www.nrs.fs.fed.us/ef/marcell/	http://www.fs.fed.us/psw/ef/san_dimas	http://www.srs.fs.usda.gov/charleston/santee

[a]amsl = above mean sea level.
[b]Water years taken as starting in May and ending in April.
[c]PET estimated from Thornthwaite (1948) method
[d]PET estimated from Hamon (1963) method with corrections.
[e]PET estimated from evaporation pan (Patric and Goswami, 1968).
[f]VSA = Variable Source Area

flood modulation due to surface storage and/or highly permeable soils. If the lower end of a curve is flat, the watershed sustains baseflow during dry periods, through release from a stored water source (e.g. groundwater), whereas a steep slope indicates a tendency for streams to dry up due to seasonality in precipitation and/or evapotranspiration and relative lack of storage. Because FDCs depict these streamflow attributes, they are important for water resources planning, especially for water uses that are influenced by extreme high and low flows. We also use the ratio of the 90th and 50th percentile daily flow (Q_{90}/Q_{50}) as an index of baseflow to assess its pattern among the watersheds, with higher values representing relatively higher baseflow or more stable flow.

Long-term (>25 years) mean daily flows are averaged for each month to characterize seasonal variability within and among sites, which assists in identifying controlling factors that cannot otherwise be captured by FDCs. The dryness index (DI; ratio of mean annual potential evapotranspiration to precipitation) is used as an indicator of energy-limited (DI < 1) versus moisture-limited (DI > 1) watersheds (e.g. Creed *et al.*, 2014). In the next section, we describe the setting and environmental features of each of the ten USDA-EFR reference watersheds evaluated. Key characteristics are compared in Table 14.1.

14.2 Site Description

14.2.1 Caribou-Poker Creek Research Watershed (CPCRW), reference sub-watershed C2, Alaska

The CPCRW is located near Chatanika in interior Alaska (Fig. 14.1) and is representative of the northern boreal forest. The 520 ha C2 reference watershed is isolated and free of any human intervention. The vegetation in CPCRW is dominated by birch and aspen on the south-facing slopes and black spruce forests on the north-facing slopes. The climate is typically continental with warm summers and cold winters.

The CPCRW is unique among the watersheds in this cross-site comparison because it is underlain by discontinuous permafrost. The permafrost distribution within the watershed exerts a strong influence over hydrological patterns (Jones and Rinehart, 2010). Studies show that as the areal extent of permafrost increases, peak discharge increases, baseflow decreases and response to precipitation events increases (Bolton *et al.*, 2004). The C2 watershed was chosen as a reference watershed because it is underlain by only about 3% permafrost compared with the adjacent C3 and C4 watersheds which are underlain by 53% and 19%, respectively.

Total mean precipitation in the C2 watershed is 412 mm, with mean snowfall and rainfall being 130 mm and 280 mm, respectively (Bolton *et al.*, 2004). Annual maximum snow depth averages 750 mm with a snow water equivalent of 110 mm. Of the total precipitation, nearly 20% becomes streamflow while evapotranspiration makes up over 75% (Bolton *et al.*, 2004). About 35% of the total precipitation falls as snow between October and April. Snowfall peaks around January while rainfall peaks around July. The spatial distribution of rainfall amount is influenced by elevation.

The relatively flat FDC for the C2 watershed (Plates 11 and 12, Table 14.2) may be attributed to the relatively well-drained soils that allow infiltration to deeper subsurface reservoirs. Runoff is generated only when the infiltration capacity is met. Streamflow in the watershed is generated by shallow subsurface storm runoff from permafrost-dominated areas, but steady groundwater baseflows with the highest Q_{90}/Q_{50} of all the sites (Table 14.2) are produced from permafrost-free areas such as C2. Spring snowmelt is usually the major hydrological event of the year, although the annual peak flow usually occurs during summer rainstorm events, as the highest rainfall intensities are greater than the maximum snowmelt rate on a daily timescale (Kane and Hinzman, 2004). It may be noted from Fig. 14.2 that the mean monthly streamflow of C2 is relatively even over the months of April through October. During winter the gauges are mostly frozen and any flow is hardly recorded, except for relatively warm temperatures. Although rainfall peaks around July, there is an increase in mean flow from July to September due to an increase in baseflow.

Table 14.2. Daily flow values for various percentage time exceedance of the flow at the ten study sites.

Watershed #/name/location	No. of daily records	Daily flow, Q (mm), for percentiles									
		0.1	1	5	10	25	50	75	90	95	Q_{90}/Q_{50}
C2/CPCRW/Alaska	4,058	3.5	2.3	1.6	1.2	0.78	0.51	0.32	0.22	0.17	0.43
NF/CCEW/California	7,671	68.0	25.3	8.9	4.5	1.13	0.27	0.08	0.04	0.03	0.15
WS18/CHL/North Carolina	27,482	22.6	11.8	7.0	5.5	3.70	2.04	1.06	0.62	0.47	0.30
WS4/FnEF/West Virginia	21,430	34.6	15.4	6.8	4.4	2.00	0.78	0.14	0.02	0.00	0.026
ESL/FrEF/Colorado	11,687	14.5	9.6	7.1	5.4	2.79	1.16	0.63	0.41	0.26	0.35
WS02/HJAEF/Oregon	22,280	66.6	29.1	15.1	9.3	4.01	1.43	0.38	0.18	0.13	0.126
WS3/HBEF/New Hampshire	20,181	51.4	24.2	9.8	5.5	2.33	0.92	0.31	0.06	0.03	0.067
S2/MEF/Minnesota	19,723	14.1	5.7	2.4	1.3	0.30	0.02	0.00	0.00	0.00	0.00
Bell 3/SDEF/Californa	18,518	30.8	4.7	1.0	0.4	0.12	0.01	0.00	0.00	0.00	0.00
WS80/SEF/South Carolina	11,256	41.7	16.8	4.2	2.1	0.42	0.03	0.00	0.00	0.00	0.00

CPCRW = Caribou-Poker Creek Research Watershed; CCEW = Caspar Creek Experimental Watershed; CHL = Coweeta Hydrologic Laboratory; FnEF = Fernow Experimental Forest; FrEF = Fraser Experimental Forest; HJAEF = H.J. Andrews Experimental Forest; HBEF = Hubbard Brook Experimental Forest; MEF = Marcell Experimental Forest; SDEF = San Dimas Experimental Forest; SEF = Santee Experimental Forest.

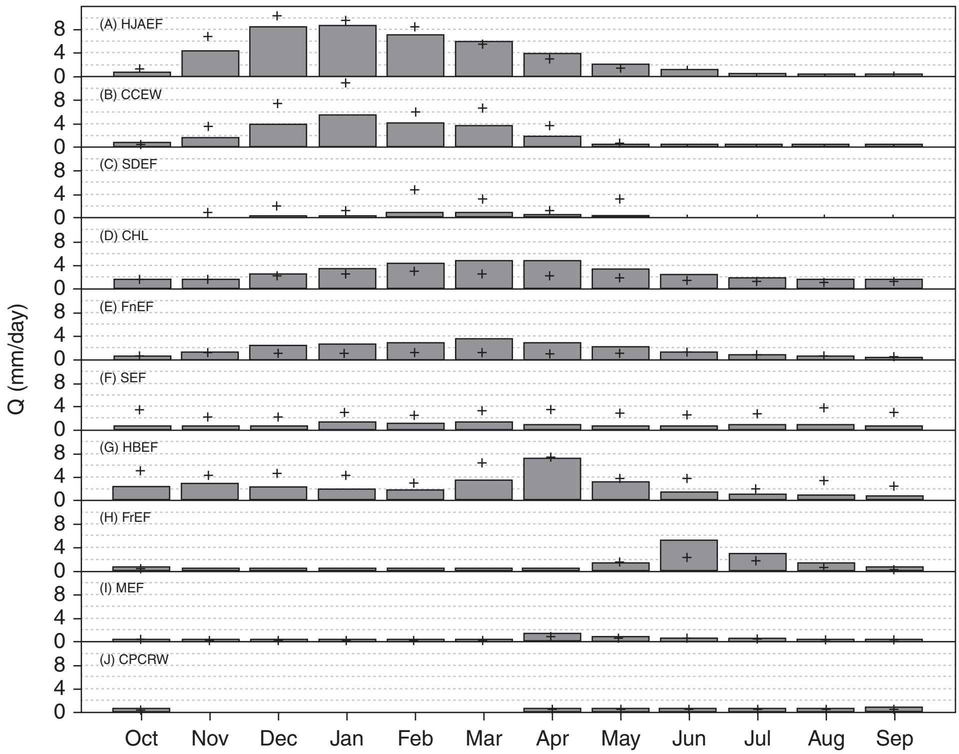

Fig. 14.2. Monthly mean daily streamflow, *Q*, averaged over the record period for each month, arranged by climate and region. '+' sign indicates standard deviation (SD) of daily flow by month. FrEF mean flow was estimated by regression of baseflow for November to May and SDs are not presented. Sample size was insufficient for flow at CPCRW for the months of November to May (HJAEF = H.J. Andrews Experimental Forest; CCEW = Caspar Creek Experimental Watershed; SDEF = San Dimas Experimental Forest; CHL = Coweeta Hydrologic Laboratory; FnEF = Fernow Experimental Forest; SEF = Santee Experimental Forest; HBEF = Hubbard Brook Experimental Forest; FrEF = Fraser Experimental Forest; MEF = Marcell Experimental Forest; CPCRW = Caribou-Poker Creek Research Watershed).

14.2.2 Caspar Creek Experimental Watershed (CCEW), reference watershed North Fork (NF), California

Located in a coast redwood and Douglas fir forest on the Jackson Demonstration State Forest in north-western California (Fig. 14.1), the CCEW hosts research designed to evaluate the effects of timber management on watershed processes. Initially, the entire 473 ha NF watershed served as the reference watershed, but after portions were logged in 1985, three NF sub-watersheds (16 to 39 ha) were designated as long-term reference watersheds. Bedrock is marine sandstone and shale of the Franciscan Complex. Most soils are 1–2 m deep loams and clay-loams and underlain by saprolite at depths of 3–8 m near ridgetops. Only about one-fifth of the 4.6 km/km^2 drainage density supports perennial streamflow. Timber production has been the major land use, and evidence of 19th century logging and the impacts of this legacy persist.

Snow is hydrologically insignificant and 95% of rainfall occurs in October–April. Fog occurs on about one-third of days in June–September, reducing summer transpiration (Keppeler, 2007). The marine influence ensures that summer air temperatures rarely exceed 20°C and winter minimums seldom drop below 0°C.

Stream runoff is about half of the average annual rainfall (Reid and Lewis, 2009). Transpiration and canopy evaporation account for nearly equal portions of the remainder (Fig. 9.1, Chapter 9, this volume). Actual evapotranspiration is limited by soil moisture deficits in May–September. Analysis of climate-related trends suggests that autumn rainfall and streamflow have declined, but with no change in annual totals.

The FDC for CCEW spans a wide range of streamflow compared with most of the other USDA-EFR sites (Plates 11 and 12) due to the strong seasonal pattern of large, episodic winter rain events that typically produce multiple, short-duration peak flows while extended summer droughts result in a long, slow recession for about half the year (Fig. 14.2). Summer streamflow is generated primarily from groundwater, and by autumn about 300 mm of precipitation is needed to mitigate moisture deficits sufficiently to generate stormflow. Stormflow (total flow based on difference between initial discharge at start of runoff and the discharge at 3 days following the cessation of the rainfall event) comprises about two-thirds of annual runoff (Reid and Lewis, 2009). Infiltration is rapid on uncompacted soils and vertical throughflow dominates near the surface. A deeper clay-rich argillic horizon can impede downward flow and generate lateral subsurface flow, although preferential flow through interconnected soil macropores limits pore-pressure increases and the extent of this perched flow. Perennial and intermittent soil pipes occur in the upper 2 m of the regolith and are frequently encountered near channel heads. When transient groundwater tables rise to the elevation of these pipes, they rapidly transmit subsurface flow to channels, mitigating pore-pressure increases upslope (Keppeler and Brown, 1998). Saturation-excess overland (return) flow is limited, but can occur on valley bottoms during large storms.

14.2.3 Coweeta Hydrologic Laboratory (CHL), reference watershed WS18, North Carolina

The CHL is located in western North Carolina (Fig. 14.1) and is representative of southern Appalachian mixed deciduous hardwoods. The 13 ha WS18 watershed was last selectively harvested in the early 1920s prior to the establishment of the CHL (Douglass and Hoover, 1988). Although the watershed has not been actively managed for more than 80 years, there have been several natural disturbances that have altered forest structure and species composition, including Chestnut blight fungus (*Endothia parasitica*) in the 1920s–1930s, drought in the 1980s and 2000s, Hurricane Opal in 1995, and hemlock woolly adelgid (*Adelges tsugae*) defoliation from 2002 to the present (Boring *et al*., 2014).

Precipitation in WS18 averages 2010 mm/year; it is highest in the late winter months and lowest in the autumn, although a disproportionate amount of large events associated with tropical storms occurs during this season. Less than 10% of precipitation occurs as snow. The variability in precipitation has been increasing over time resulting in more frequent extremely wet years and extremely dry years, while annual average air temperature has been increasing by 0.5°C/decade since 1981 (Laseter *et al*., 2012).

Annual precipitation in WS18 is approximately equally partitioned into streamflow (49.6%) and evapotranspiration (50.4%). During the growing season, transpiration accounts for 55% of total evapotranspiration with evaporation from canopy interception making up the balance, approximately 15% of precipitation (Ford *et al*., 2011). Streamflow is typically highest in March–April and lowest in September–October but never ceases, even during extreme drought. Seasonal patterns in streamflow reflect the combined effects of the seasonality in precipitation and evapotranspiration (Fig. 14.2).

Baseflows are relatively high, producing the third largest Q_{90}/Q_{50} ratio among sites (Table 14.2). Baseflows are sustained by lateral movement of water through deep unsaturated soil (Fig. 9.1, Chapter 9, this volume), driven by large hydraulic gradients induced by steep slopes (Hewlett and Hibbert, 1963). On average, approximately 5% of annual precipitation (9% of annual streamflow) is discharged as stormflow (Swift *et al*., 1988). Stormflow originates primarily from small portions of the watershed located adjacent to the stream in coves and in riparian zones where the water table may be near the surface (Hewlett and Hibbert, 1967). Shallow lateral subsurface discharge from upslope landscape positions to streams can also contribute to stormflow where large soil macropores exist. Overland flow is extremely rare or non-existent because of the presence of well-developed forest floors and subsurface macropores.

14.2.4 Fernow Experimental Forest (FnEF), reference watershed WS4, West Virginia

The FnEF is located in eastern West Virginia (Fig. 14.1) and is representative of the 'unmanaged' forests of the central Appalachian region. The 39 ha WS4 watershed is forested with an approximately 100-year-old stand of mixed deciduous hardwoods. The bedrock is acidic sandstone and shale. Depth to bedrock is generally less than 1 m and the topography is steep.

Precipitation is distributed evenly throughout the year and averages 1458 mm. Although snow is common in winter, snowpack generally lasts no more than a few weeks; snow contributes approximately 14% on average of precipitation (Adams *et al*., 1994). Large rainfall events can occur during extra-tropical hurricanes in the summer and autumn, but about half of the largest storms have occurred during the dormant season (1 November–30 April), when streams are most responsive to rainfall because evapotranspiration losses are low (Fig. 14.2).

The stream channel is intermittent near the top of the watershed. Streamflow may cease during the late summer and early autumn (about 10% of daily flows), in response to high evapotranspirative demand and low precipitation. Although baseflow contributes relatively little to Q_{90}/Q_{50} (Table 14.2), it dominates stream discharge in WS4. Most discharge occurs during the dormant season (Fig. 14.2) due to greater precipitation and decreased evapotranspirative demand from deciduous forests. Baseflow is sustained by lateral subsurface flow to channels; DeWalle *et al*. (1997) characterized the mean transit time for baseflow on WS4 as 1.4–1.6 years, which suggests a dominance of slow movement through the soil matrix.

The water balance on WS4 was well quantified by Patric (1973) with runoff accounting for about 40% of precipitation, 27% of the balance being lost through transpiration and about 16% to canopy evaporation. Seasonal differences in losses from canopy interception due to leaf development and leaf drop were detected.

Stormflow discharge is fairly flashy (Plates 11 and 12), with the storm hydrograph responding rapidly to storm precipitation inputs and then returning quickly to baseflow conditions, and streamflow generation occurs via saturation excess flow. Stormflow discharge typically occurs less than 15% of the time. There is little to no infiltration-excess overland flow even during the largest storms because of the high infiltration capacity of an intact forest floor.

14.2.5 Fraser Experimental Forest (FrEF), reference watershed East St Louis (ESL), Colorado

The FrEF is located in the Rocky Mountain cordillera of Colorado (Fig. 14.1) and is representative of subalpine watersheds over a large portion of the central Rockies. It spans the subalpine to alpine zone; a zone that is characterized by relatively low temperatures and moderate precipitation (Love, 1960). The area is dominated by Engelmann spruce and subalpine fir on higher-elevation and shaded slopes, lodgepole pine on lower-elevation sunny slopes and alpine tundra above the treeline. The 803 ha ESL watershed has received no significant treatment in over 90 years (Retzer, 1962).

Precipitation is dominated heavily by snowfall (about 75%) from October through May (Alexander *et al*., 1985) and runoff is dominated by snowmelt (about 90%) from May through August (Fig. 14.2). Significant summertime convective rainfall events may also temporarily increase flow. The main stem is perennial but baseflow is low, stable and unmeasured during the winter months due to logistical difficulties of stream measurements in winter.

The runoff coefficient for annual flow is about 45% with significant wintertime sublimation losses from the canopy and summertime evapotranspiration. Summertime rainfall is primarily used on site by vegetation, with high evaporative losses due to dry air masses and wind.

High-elevation stream reaches are intermittent with spring and summertime flows fed by snowmelt (Fig. 14.2). The hydrological regime is dominated by a typical seasonal snowmelt hydrograph with a rapid rising limb in May and June, followed by a long recession, returning to baseflow (second largest Q_{90}/Q_{50}, Table 14.2) in August (Alexander *et al.*, 1985; Troendle and King, 1985). Extensive spring networks feed the drainage systems as the annual snowmelt pulse moves through the basin (Retzer, 1962). Rainfall events punctuate the snowmelt hydrograph, but contribute insignificant amounts to the annual runoff. Infiltration-excess overland flow is rare, but may occur under the snowpack during the melt season when frozen ground impedes infiltration. Saturation-excess overland flow is extremely rare as infiltration rates for the porous soils and glacial till typically exceed maximum rainfall and snowmelt rates (Retzer, 1962).

The ESL represents the highest elevation range, largest snowpack and largest watershed of this cross-site comparison. Maximum snowmelt rates are limited by incoming energy and can never reach extreme rainfall rates. Rain-on-snow flood events can alter flow statistics, but are rare in this portion of the Rockies. The relatively large size of the basin also reduces flashy response or high runoff per unit area observed in smaller basins.

14.2.6 H.J. Andrews Experimental Forest (HJAEF), reference watershed WS02, Oregon

The HJAEF is located in the western Cascade Mountains of central Oregon (Fig. 14.1) and is representative of Pacific Northwest moist conifer forests. Watershed 2 (WS02) is 60 ha and the geology is dominated by bedrock of volcanic origin. Stream channels are steep and confined with unsorted sediment dominated by cobbles and boulders, with patches of silt and exposed bedrock. Shallow hillslope soils (generally less than 1 m deep) are loam and clay loam. Stone content ranges from 35 to 80%, increasing on south-facing slopes. The steep hillslopes in WS02 are dominated by 500- to 550-year-old Douglas fir (*Pseudotsuga mensiesii*) forests with western hemlock (*Tsuga heterophylla*) and western red cedar (*Thuja plicata*) (Rothacher *et al.*, 1967). The canopy is greater than 60 m tall. The climate is continental with cold winters and cool, short, dry summers.

Annual precipitation averages 2300 mm, falling primarily as rain between November and April and with occasional snow at higher elevations. Soil temperatures remain above freezing. The annual hydrograph in WS02 has a strong seasonal pattern with a high winter baseflow and frequent autumn, winter and spring stormflows in contrast to very low flows in summer (Fig. 14.3).

Approximately 57% of the precipitation is streamflow (Post and Jones, 2001). Baseflow accounts for only 43% of the discharge (Q_{90}/Q_{50} = 0.126) (Table 14.2) whereas quickflow comprises the remainder (Fig. 9.1, Chapter 9, this volume). McGuire *et al.* (2005) estimated that mean baseflow residence time for WS02, based on δO^{18} of water, was approximately 2.2 years. They suggested that topography and steepness may be exerting greater control on residence times than watershed area. Although there are no detectable trends in streamflow from 1987 to 2007, in more recent time periods (1996–2007) slight decreasing trends have been observed (Argerich *et al.*, 2013).

The relatively steep FDC for WS02 (Plates 11 and 12) has been attributed to highly permeable soils and strong seasonal precipitation patterns. Fast percolation rates, typically greater than 0.12 m/h, are influenced by high stone content and large pore spaces (Rothacher *et al.*, 1967). These characteristics also lead to substantial hyporheic flows lateral to and beneath the streams (Kasahara and Wondzell, 2003).

14.2.7 Hubbard Brook Experimental Forest (HBEF), reference watershed W3, New Hampshire

The HBEF is located in New Hampshire (Fig. 14.1) and is representative of mature northern hardwood stands. Vegetation at W3 is composed mainly of sugar maple (*Acer saccharum*), American beech (*Fagus grandifolia*) and yellow birch (*Betula alleghaniensis*). The 42 ha watershed is mostly second growth and much of the HBEF was harvested in the 1910s (Table 14.1). Additional salvage harvesting occurred at the HBEF following the Great New England Hurricane of 1938. More recently, trees incurred some damage during the North American Ice Storm of 1998, with no apparent impact on annual runoff.

The climate at the HBEF is cool and humid. On average, W3 receives 1350 mm of precipitation annually, which is distributed evenly throughout the year. Precipitation has increased by 25% during the record period, which is consistent with broader regional trends (Brown *et al*., 2010). Approximately one-third of precipitation falls as snow (Fig. 9.1, Chapter 9, this volume) and a snowpack generally persists from late December until mid-April. Soil frost forms during winter two out of every three years with an average annual maximum depth of 6 cm.

The annual hydrograph shows a strong seasonal pattern with a peak during snowmelt runoff. Despite the higher flow during spring, floods can occur at any time of year when soil water deficits are reduced (Fig. 14.2). An increasing trend in precipitation has resulted in increasing trends in the magnitude of both low and high streamflows (Campbell *et al*., 2011).

Approximately 64% of the precipitation that falls on the watershed becomes streamflow, with evapotranspiration comprising the remainder. Slight, but statistically significant declines in evapotranspiration have occurred in W3 (14% over 56 years) for reasons that are unknown. This decline appears to be due to local influences since similar trends are not consistently found at a larger regional scale.

The relatively steep FCD for W3 (Plates 11 and 12) has traditionally been attributed to coarse, well-drained soils and mountainous topography that produce a flashy runoff response. Overland flow is also minimal because of the high infiltration capacity of the forest floor. In recent years, a more complete understanding of complex flow generation processes at the site has emerged. Data from a network of wells in W3 have revealed an intermittent, discontinuous water table (Detty and McGuire, 2010a; Gannon *et al*., 2014; Gillin *et al*., 2015). Stormflow generation is the result of lateral subsurface flow in the solum. Under some soil moisture conditions, small changes in groundwater can produce large changes in runoff, suggesting a threshold response that is related to flowpaths and soil transmissivity (Detty and McGuire, 2010b; Gannon *et al*., 2014). During low flows, only the near-stream zone is consistently hydrologically connected to the stream network. As the watershed wets up, more distal, previously isolated portions of the water table become hydrologically connected.

14.2.8 Marcell Experimental Forest (MEF), reference watershed S2, Minnesota

The MEF is located along the southern fringe of the boreal biome, in northern Minnesota (Fig. 14.1). The landscape includes uplands, peatlands, lakes and streams. Unlike mountainous research watersheds, streamflow typically is not bedrock controlled in the western lakes section where outwash sands, some >50 m deep, form large aquifers (Verry *et al*., 2011). Aquifer–peatland connectivity varies between two peatland types: bogs and fens (Bay, 1967). In watersheds with either type, streamflow may originate from precipitation and flow along near-surface and shallow surface flowpaths in upland mineral soils (Verry *et al*., 2011). Bog watersheds may be perched due to loamy clay tills that retard the vertical flow of water from soils to the outwash aquifer (Verry *et al*., 2011). In fen watersheds, most streamflow, which may exceed streamflow from bogs by orders of magnitude during low flow, originates as discharge from aquifers and is perennial (Bay, 1967).

The 10 ha S2 study watershed, with a bog (33% of the area), has low topographic relief (Table 14.1) with upland mineral soils that drain through peatland margins to an intermittent stream. Eleven to 33% of annual precipitation (456–981 mm) occurs as streamflow and 5–17% recharges the underlying aquifer (Nichols and Verry, 2001) (Fig. 9.1, Chapter 9, this volume). Calculated evapotranspiration (precipitation – streamflow – recharge) has been 372–605 mm/year. Nine of the ten highest daily streamflows have occurred during rainfall–runoff events, not snowmelt or rain-on-snow events. Periods of no streamflow occur during any month and there has been no flow during 38% of the record (Plates 11 and 12), consistent with the zero value of Q_{90}/Q_{50} (Table 14.2).

Although most of the S2 area is uplands, most of the annual water budget (58%) comes from direct precipitation on the peatland (Verry *et al.*, 2011). If the water table is >5–10 cm below the peatland surface, streamflow ceases and that storage must be replenished before resumption. Rainfall amount during summer exceeds snow water equivalents during winter and stormflows recess rapidly to no flow due to evapotranspiration. Melt from snow accumulation (November/December to March/April) results in several weeks of high flows (Sebestyen *et al.*, 2011) (Fig. 14.2). Winter and spring frost in upland soils, exceeding 50 cm, prevents infiltration (Verry *et al.*, 2011). Snowmelt waters flow overland until soils thaw in the spring, after which flow mostly occurs in the shallow subsurface through sandy loams above loamy clay horizons (Verry *et al.*, 2011). Subsurface flow may persist for weeks until the upland deciduous forest begins transpiring. During large summer rainfall events, subsurface flow may last for several hours, but rarely longer.

14.2.9 San Dimas Experimental Forest (SDEF), reference watershed Bell 3, California

The 25 ha watershed at SDEF is located in southern California (Fig. 14.1) and is representative of the chaparral forests of the US Southwest. Chaparral forest is a dense, drought-tolerant shrubland with a closed canopy some 3–5 m in height. Chaparral is a fire-prone ecosystem and wildfires have burned the SDEF about every 40 years.

Regional hydrology is controlled by climate and geology: cool, wet winters followed by long summer droughts; and ongoing tectonic uplift that has produced steep topography and exposed fractured crystalline basement rocks that weather to thin, coarse-textured, azonal soils (Dunn *et al.*, 1988) (Table 14.1). Precipitation falls almost exclusively as rain from winter frontal storms and rare summer thunderstorms. Nearly 90% of the annual rainfall occurs between December and April with the most runoff in February (Fig. 14.2).

Streamflow accounts for only roughly 11% of the rainfall, with the remainder apportioned to evapotranspiration and groundwater recharge. Groundwater dynamics on the SDEF are virtually unknown, rendering the closure of any water balance exercise moot. However, groundwater recharge is potentially large through the fractured substrate, reducing any calculated value of actual evapotranspiration. Soil moisture is at or below the wilting point by the end of the summer and the drought-adapted plants likely get their water from fractures in the bedrock.

Stream runoff is generated by saturation excess flow in riparian zones, presumably as shallow throughflow moves laterally through the coarse soil mantle (Fig. 9.1, Chapter 9, this volume). Infiltration-excess overland flow on hillside slopes is rare and occurs only during the most intense rainstorms, reflecting the high infiltration rates of the soil and percolation into bedrock. However, after wildfire, with the combustion of the canopy and surface litter layer as well as changes in soil properties (bulk density and water repellency), hillslope hydrology shifts to pervasive overland flow after saturation of the very thin surface wettable layer (Rice, 1974; DeBano, 1981). Water that formerly slowly flowed by subsurface pathways now moves quickly into the stream channels, increasing runoff for comparable storms by up to four orders of magnitude over pre-fire levels (Wohlgemuth, 2016). The effects of fire on the forest hydrology can persist for several years.

14.2.10 Santee Experimental Forest (SEF), reference watershed WS80, South Carolina

The SEF is located in eastern South Carolina (Fig. 14.1) and is representative of the subtropical coastal watersheds throughout much of the US Southeast, with hot and humid summers and moderate winter seasons. The 155 ha WS80 watershed is covered with a pine/mixed hardwood forest (Table 14.1), which has been undisturbed by management activities since 1936, but was heavily affected by Hurricane Hugo in 1989 that damaged >80% of the forest canopy (Hook *et al.*, 1991).

Seasonally, the winter is generally wet with low-intensity, long-duration rain events and rare snowfall. Summer is characterized by short-duration, high-intensity storm events and tropical depression storms are common. The seasonal runoff response to rain events is shown in Fig. 14.2.

Approximately 22%, on average, of annual precipitation becomes runoff (Amatya *et al*., 2006), resulting in about 78% evapotranspiration, assuming negligible seepage (Fig. 9.1, Chapter 9, this volume). Approximately 60% of the runoff is contributed by shallow surface or runoff/rainwater, the rest by subsurface flow (Epps *et al*., 2013).

Based on the FDC analysis this watershed produces flow only 56.3% of the time and hence has a zero value of Q_{90}/Q_{50} (Plates 11 and 12, Table 14.2). The principal flow generation mechanism is driven by the shallow water table (Fig. 9.1, Chapter 9, this volume) (Harder *et al*., 2007; Epps *et al*., 2013), controlled primarily by rainfall and evapotranspiration, and minimally by deeper groundwater underlain by Santee Limestone approximately 20 m below the ground surface. The formation of an argillic horizon with poorly drained clayey subsoil provides a dynamic shallow groundwater table that has a complex non-linear relationship with streamflow (Harder *et al*., 2007). Saturation-excess surface and shallow subsurface runoff with rapid lateral transfers within the highly permeable upper soil layer may occur along reaches with flat topography. Surface depressional storage was shown to affect the surface runoff rate (Amoah *et al*., 2012). Runoff and peak flow at this watershed are dependent on both rainfall amount and intensity, as well as antecedent conditions reflected by initial water table positions (Epps *et al*., 2013).

A key observation from WS80 is the reversal of the flow relationship between this and the treatment watershed, compared with the earlier calibration period, for a decade beginning three years after Hurricane Hugo severely damaged vegetation on both watersheds. As a result reduced evapotranspiration in selected hurricane-affected vegetation on the reference watershed enhanced its streamflow (Jayakaran *et al*., 2014). Long-term data also indicate rising air temperature and increasing frequency of large storms (Dai *et al*., 2013).

14.3 Discussion of Hydrological Processes

14.3.1 Flow duration curves

FrEF and CPCRW host the largest reference watersheds among our study sites (Table 14.1). FrEF has the highest elevation range, deepest snowpack and the largest drainage area. These factors, combined with the snowmelt-driven hydrological regime, explain the somewhat different behaviour in flow duration with higher flow values for FrEF than for CPCRW (Plates 11 and 12, Table 14.2). The muted high flows, with their greater influence at CPCRW potentially due to its relatively well-drained soil conditions (see Section 14.2.1 above), are most likely attributed to the large size of these watersheds. However, this does not hold true for CCEW which, although comparable in size to CPCRW (Table 14.1), has a steep FDC for low exceedance, perhaps due to its much larger seasonal precipitation, deep clay horizon and soil pipes that contribute to a rapid runoff response (see Section 14.2.2 above).

In comparison, SDEF has the second smallest reference watershed and forth steepest watershed examined (Table 14.1). As a result, its FDC shows very flashy storm responses followed by long, declining flows that eventually are zero for 47.5% of the record. Similarly, MEF, characterized by deep peat and possibly high storage capacity, and SEF with shallow sandy clay loam soils generate no surface flow for 44% of their periods of record, with $Q_{90}/Q_{50} = 0$ for all three sites (Table 14.2). Although SEF is the lowest gradient watershed, the high flow range that occurs for less than 1% of the time is greater than at most of the other sites, except for HBEF, HJAEF and CCEW. The highest flows at this site result from storm runoff from saturated clay-rich soils (Epps *et al*., 2013; Griffin *et al*., 2014).

Along the Coast Range of the western USA, HJAEF and CCEW have FDCs that are similar in shape, likely related to seasonal climatic patterns. The HJAEF has the third steepest basin slope after CHL and CCEW (Table 14.1) but the highest FDC slope for low flows occurring more than 0.2 to 30% of the time, above which the CHL has the highest low flow (Plates 11 and 12). Although WS02 at HJAEF is smaller than the watershed at CCEW (Table 14.1), it generally sustains higher flows, except at the lowest exceedance frequencies, likely because it receives 1.75 times more precipitation than the CCEW. Both of these western watersheds have similar forest species and leaf area index (LAI) (Table 14.1) as well as frequent large storms in winter and dry summers. Weiler and McDonnell (2004) suggest additional factors including lateral soil conductivity and drainable porosity may explain variability in streamflow response, specifically at HJAEF.

CHL has the steepest basin slope (52%) of all the watersheds in this analysis and a 95th percentile flow (Q_{95}) of 0.47 mm/day, which is the largest of all the sites (Table 14.2). Of the three sites in the Appalachian Mountains (i.e. CHL, FnEF and HBEF), CHL also has the smallest drainage area and is more southerly than FnEF and HBEF (Table 14.1). Interestingly, this reference watershed also has the highest Q_{90}/Q_{50} values (indicative of sustained baseflow) and lowest flow values for the higher flow ranges ($Q_{0.1}$ or lower exceedance) but has equal or higher flows at and above Q_{25} compared with FnEF or HBEF (Table 14.2). The higher flow in the lower exceedance range in the more northern HBEF site could be partially attributed to snowmelt and the higher flow in lower exceedance range at the CHL site is likely due to sustained baseflows caused by high storage of deep soils (Table 14.1).

Although on opposite coasts, the 61 ha HJAEF site yields consistently higher percentile flows (Table 14.2) compared with the 42.4 ha HBEF site at almost the same latitude, similar elevations, potential evapotranspiration, and surface and subsurface flow generation mechanisms (Table 14.1). The exception is the extreme high end of discharges at or below 0.01% exceedance when both exhibit a similar pattern (Plates 11 and 12), which is attributed to the HJAEF having higher slope and 41% higher precipitation than the HBEF. In their analysis of threshold hydrological response across northern catchments, Ali *et al.* (2015) found some similarities in rainfall- and snowmelt-driven events between these two watersheds.

14.3.2 Long-term mean daily flow

Figure 14.2 (plots A–C) shows long-term mean daily flow by month for west-coast watersheds which all have strongly seasonal rainfall. Oregon's HJAEF (plot A) has the greatest monthly flows, with a longer winter rainy season than the more southerly sites. In California, coastal CCEW (plot B) reflects the transition from the wetter north-west to the arid Mediterranean climate of SDEF (plot C). These three western sites show highly variable winter flow patterns due to the episodic nature of the Pacific frontal systems with increased coefficient of variation further south where large winter storms are less frequent. These patterns are also consistent with the relative variability defined by the upper and lower exceedance percentiles of the FDCs (Plates 11 and 12, Table 14.2).

Similarly, the east-coast watersheds in Fig. 14.2 (plots D–G) range from high mean flow in the winter to low flow in the summer and early autumn, with the exception of HBEF (plot G). CHL (plot D) shows a smooth annual hydrograph that peaks in late spring following the seasonal rainfall pattern. FnEF (plot E) and SEF (plot F) have similar mean annual precipitation, but the SEF produces less than half of the runoff generated at FnEF, primarily due to higher potential evapotranspiration (Table 14.1). The seasonal signal for the FnEF and CHL reflects their inland locations and a more pronounced dormant season relative to SEF. Both CHL and FnEF show relatively little streamflow variability due to relatively consistent precipitation with little variance. The relatively high streamflow variability at the SEF results from a dynamic water table regulated by coastal climate and shallow clayey argillic horizon. HBEF (plot G) is well north of the other east-coast basins, putting it in a location where snow plays a greater role in the hydrological regime. It is the only watershed in the study that shows a significant double peak in annual flow: a rainfall peak in November and a snowmelt or rain-on-snow peak in April.

Snowmelt and continentality have a dominant influence on annual water budgets in the last three study areas: FrEF (plot H), MEF (plot I) and CPCRW (plot J) (Fig. 14.2). FrEF receives most of its precipitation in the form of wintertime snow. The CPCRW (plot J) represents an extreme in almost every metric used (Table 14.1) including the annual precipitation and runoff. All of the snowmelt-dominated watersheds show lower relative variance in flow because the peak flows are regulated by both the amount of snow and the maximum amount of energy available to melt snow, with the occasional exception at the MEF where some peak flows occur during rain-on-snow events. In general, higher mean monthly flows are observed in basins close to coastal moisture sources or at lower latitudes, although there are exceptions (SDEF, SEF). Higher variances are also observed near coasts, where large, episodic rainfall events are more influential. Snowmelt processes reduce variance (FrEF, MEF and CPCRW), while inland watersheds also exhibit less variability in daily mean flows (FnEF and CHL).

14.3.3 Other watershed characteristics affecting hydrology

Data from these ten sites show that none of the parameters in Table 14.1 (temperature, potential evapotranspiration, drainage area, altitude, latitude) has a significant influence on annual streamflow, except for annual precipitation, which is found to be a strong driver ($R^2 = 0.85$), as expected. However, annual evapotranspiration, calculated as the difference between precipitation and streamflow (i.e. not considering groundwater recharge), correlates well ($R^2 = 0.72$) with an independent estimate of potential evapotranspiration, and also with temperature ($R^2 = 0.76$) and latitude (inversely, $R^2 = 0.53$), as expected. Another interesting finding is that sites (CPCRW, MEF, SDEF and SEF) with DI values higher than 0.71 closer to soil moisture limited have a much lower (0.12–0.22) average runoff coefficient (streamflow/precipitation) than the remaining energy-limited sites (0.44–0.64) which have a DI < 0.50 (Table 14.1). Although most of the site characteristics for the HBEF and HJAEF are similar, except for precipitation which is higher at the HJAEF, the streamflow as a percentage of precipitation for the HJAEF is actually lower than that of the HBEF. This is possibly due to the higher evapotranspiration of its conifer forest, with its LAI almost twice that of the northern hardwood forest at the HBEF site. However, other factors such as geology and lithology besides the evapotranspiration might also be influencing losses. FrEF receives similar precipitation to SDEF and MEF, but has two to four times the annual streamflow because of much lower potential evapotranspiration as well as runoff occurring in a relatively steep basin, over a concentrated period, when a significant portion of the vegetation is dormant. However, some seepage to a regional aquifer at the MEF and possible groundwater recharge at the SDEF are also factors in their lower flow.

14.3.4 Implications of hydrological processes

Improved understanding of runoff generation and flowpaths helps land managers identify hydrologically connected areas that contribute to streamflow and pollutant discharge. The synthesis of runoff patterns across sites (Plates 11 and 12, Fig. 14.2) is important for identifying relationships between streamflow and nutrients that aid in developing load duration curves used to establish water quality standards (Argerich *et al*., 2013). This important information is being used to assess the impacts of forest disturbance and restoration projects, and will help to better predict hydrological and chemical responses and transport. For example, monitoring procedures developed at the CCEW site are widely used to assess sediment and pollutant loads. This information is helpful in evaluating potential timber harvest impacts and in the development of forest management regulations and best management practices (Cafferata and Reid, 2013).

Knowledge of processes derived using long-term records from these diverse watersheds (Table 14.1) enables scientists to better understand their interrelationships with climate, forest vegetation and water use, and ecosystem dynamics (Vose *et al*., 2012). For example, intensively monitored plots at CHL are providing new insights into relationships between soil moisture, carbon and nitrogen cycling, and vegetation allocation processes along topographic gradients. Furthermore, these records are being used to study hydrological recovery from disturbances such as the catastrophic mountain pine bark beetle infestation at FrEF, extreme hurricanes at SEF and historic land use at CCEW.

14.4 Summary

This cross-site comparison has used long-term hydrometeorological patterns, basin hydromorphological parameters and other attributes (Table 14.1) to compare and contrast forest hydrological processes (Fig. 9.1, Chapter 9, this volume) at ten reference watersheds in the USDA-EFR network. The response of streamflow to variation in annual precipitation magnitude, form and seasonality, and evapotranspiration at each watershed was evaluated by using daily FDCs (Plates 11 and 12), as well as the long-term mean daily flow for each month (Fig. 14.2).

Statistical results (Plates 11 and 12, Fig. 14.2 and Table 14.2) in the context of key watershed variables (Table 14.1) show that these watersheds have distinct hydrological processes and, therefore, can help frame our conceptual understanding of forest runoff processes (Fig. 9.1, Chapter 9, this volume), with precipitation as a driving variable for both high and low flows. While some seasonal flow patterns were observed among sites along the eastern and western near-coastal areas, flowpaths of rain and snowmelt water were shown to vary greatly across and within reference watersheds, potentially affecting the timing and peak of storm runoff, as illustrated by the FDCs (Plates 11 and 12, Table 14.2) and long-term monthly mean daily flows (Fig. 14.3). A DI value of about 0.50–0.70 was found to be an approximate break range for identifying sites with high runoff or low runoff, relative to the precipitation received. The analysis also revealed that larger watersheds do not necessarily yield higher baseflows and damped high flows. In addition, the presence of an argillic horizon, large topographic depressions and riparian area, preferential flowpaths, pipeflow, steep slopes and certain soil physical properties also significantly affect flowpaths, the magnitude and variation of runoff generation, and possibly the water balance (Weiler and McDonnell, 2004; Griffin *et al*., 2014; Gillin *et al*., 2015; Klaus *et al*., 2015). Furthermore, the results also demonstrate that a better hydrological understanding of low topographic relief sites such as MEF and SEF is needed because these areas are common but not well represented by EFR sites, which are mostly in mountainous terrain.

Although this comparative study helps advance our understanding of runoff generation mechanisms across these diverse watersheds, increased evidence in recent years supports a nonlinear rainfall–runoff response both on hillslopes and low-gradient coastal landscapes, highlighting the need to better quantify hydrological thresholds and understand physical controls (Spence, 2010; la Torre Torres *et al*., 2011; Epps *et al*., 2013; Ali *et al*., 2015; Klaus *et al*., 2015). Research on linkages between hydrology and soil development (e.g. Gillin *et al*., 2015), peatland watershed responses to environmental and climatic change (Kolka *et al*., 2011), rainfall–runoff relationships in chaparral vegetation, interactive effects of vegetation and stand type on streamflow (Jayakaran *et al*., 2014), hydrological processes on tidally affected riparian forested wetlands (Czwartacki, 2013), etc. is advancing in some of these watersheds. Incorporation of this new information into ecohydrological models (Dai *et al*., 2010; Amatya and Jha, 2011) will improve predictions of runoff generation and our ability to assess responses to future disturbances.

Long-term data from this spectrum of watersheds demonstrate the value of the USDA-EFR network for studies of a variety of hydrological processes and their interactions in different environments, which is not possible at individual sites or using short-term studies. This variability across sites will also be critical in future studies for process-level, statistical and modelling research relating to impacts, vulnerability and risk assessments of climate and land-use change, and forest disturbance on hydrology, biogeochemistry and water supply. These reference watersheds also continue to be important for use in paired watershed studies to evaluate effects of disturbances such as forest harvesting, prescribed burning, devegetation, changes in forest structure and species composition, fertilization and other land management practices on water yield, evapotranspiration, flowpath routing, nutrient cycling and sediment transport. Indeed, the research is being used to chart long-term effects and the data collected have been essential for cross-site syntheses (e.g. Kolka *et al*., 2011; Jones *et al*., 2012; Creed *et al*., 2014; Gottfried *et al*., 2014; Vose *et al*., 2014). However, additional studies are also warranted to examine consistency of these long-term data and results from the reference watersheds used in various hydrological analyses herein and elsewhere for their potential deviation, if any, due to unforeseen external factors including climate change (Alila *et al*., 2009; Ali *et al*., 2015).

Therefore, there is a critical need for continued monitoring of these long-term watersheds, as they are well suited for documenting and detailing baseline hydrological conditions and also serve as valuable benchmarks for addressing emerging forest and water issues of the 21st century.

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