AN ABSTRACT OF THE DISSERTATION OF

V. Cody Hale for the degree of Doctor of Philosophy in Water Resources Science presented on December 19, 2011

Title: Beyond the Paired-Catchment Approach: Isotope Tracing to Illuminate Stocks, Flows, Transit Time, and Scaling.

Abstract approved:

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Jeffrey J. McDonnell

This dissertation integrates a process-based hydrological investigation with an ongoing paired-catchment study to better understand how forest harvest impacts catchment function at multiple scales. We do this by addressing fundamental questions related to the stocks, flows and transit times of water. Isotope tracers are used within a top-down catchment intercomparison framework to investigate the role of geology in controlling streamwater mean transit time and their scaling relationships with the surrounding landscape. We found that streams draining catchments with permeable bedrock geology at the Drift Creek watershed in the Oregon Coast Range had longer mean transit times than catchments with poorly permeable bedrock at the HJ Andrews Experimental Forest in the Oregon Cascades. We also found that differences in permeability contrasts within the subsurface controlled whether mean transit time scaled with indices of catchment topography (for the poorly permeable bedrock) or with catchment area (for the permeable bedrock). We then investigated the process-reasons for the observed differences in mean transit time ranges and scaling behavior using a detailed, bottom-up approach to characterize subsurface water stores and fluxes. We found that the mean transit times in catchments underlain by permeable bedrock were influenced by multiple subsurface storage pools with different groundwater ages, whereas storage in the poorly permeable catchments was limited to the soil profile and that resulted in quick routing of excess water to the stream at the soil bedrock interface, leading to mean transit times that were closely related to flowpath lengths and gradients. Finally, we examined how and where forest trees interacted with subsurface storage during the growing season using a forest manipulation experiment, where we tested the null hypothesis that near-stream trees alone influenced daily fluctuations in streamflow. We felled trees within this zone for two 2.5 ha basins and combined this with isotopic tracing of tree xylem water to test if water sources utilized by trees actively contributed to summer streamflow. We rejected our null hypotheses and found that diel fluctuations in streamflow were not generated exclusively in the near-stream zone. We were unable to link, isotopically, the water sources trees were utilizing to water that was contributing to streamflow. Our results provide new process-insights to how water is stored, extracted, and discharged from our forested catchments in Western Oregon that will help better explain how forest removal influences streamflow across multiple scales and geological conditions.

# Beyond the Paired-Catchment Approach: Isotope Tracing to Illuminate Stocks, Flows, Transit Time, and Scaling

by

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I understand that my dissertation will become part of the permanent collection of Oregon State University libraries. My signature below authorizes release of my dissertation to any reader upon request.

V. Cody Hale, Author

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## CONTRIBUTION OF AUTHORS

Chapter 2: Jeff McDonnell provided critical reviews and editing support.

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TABLE OF CONTENTS

Page

TABLE OF CONTENTS (Continued)

Page

TABLE OF CONTENTS (Continued)

Page

LIST OF FIGURES

LIST OF FIGURES (Continued)

Figure Page

LIST OF TABLES

# Introduction

## INTRODUCTION

Streams draining headwater forested watersheds are the source of nearly two-thirds of the clean water supply in the United States [*NRC*, 2008]. They also serve a number of other critical functions, such as providing habitat for aquatic biota, processing nutrients, and transporting sediments [*Gomi et al.*, 2002]. It is therefore important that we understand if and how activities occurring within our forested watersheds impact the quantity, timing, or quality of their waters. As such, the effects of forest harvest on streamflow have been the focus of extensive research efforts over the past century [*Andreassian*, 2004; *Bosch and Hewlett*, 1982; *Stednick*, 1996]. However, the findings to date have been somewhat ambiguous [*Hibbert*, 1967] and, at times, controversial [*Beschta et al.*, 2000; *Jones and Grant*, 1996; *Thomas and Megahan*, 1998].

Hydrologic response of small streams to forest harvest has traditionally been assessed using a paired-catchment experimental design [*Bosch and Hewlett*, 1982]. This approach uses statistical relationships of hydrologic metrics developed between two or more watersheds to detect change following forest harvest [*Kovner and Evans*, 1954]. Since only streamflow (and water quality) data, collected from the catchment outlets, are used in the statistical analysis, little is learned about the process interactions between forest vegetation removal and mechanisms of hydrologic response. In addition, the uniqueness of individual catchments makes it difficult, if not impossible, to replicate treatments, effectively eliminating the ability to infer research findings to other catchments in a pure statistical sense [*Hewlett et al.*, 1969; *Kovner and Evans*, 1954; *Murtaugh*, 2000]. In his analysis of 39 paired-catchment studies, *Hibbert* [1967] was unable to provide more than broad generalizations of the typical response trajectories for hydrologic metrics (i.e. annual discharge, peak flows, low flows) following forest harvest. Even with additional studies to draw from, neither *Bosch and Hewlett* [1982; 94 studies] nor *Andreassian* [2004; 137 studies] were able to expand upon *Hibbert's* [1967] generalizations. In *Hibbert's* [1967] words, "response to treatment is highly variable and, for the most part, unpredictable".

Numerical models have the potential to be used as powerful tools for predicting hydrologic effects of land-use change [*Beschta*, 1998; *Seibert and McDonnell*, 2010; *Singh and Woolhiser*, 2002; *Zegre et al.*, 2010]. However, predictive capabilities of such models are currently limited by the lack of understanding of the dominant processes controlling hydrologic response at various scales [*Blöschl*, 2001; *Sidle*, 2006; *Tetzlaff et al.*, 2011]. A more detailed depiction of the states and fluxes that make up the catchment-scale water balance is needed to improve model accuracy. Although, much progress has been achieved in gaining such understanding at the plot and hillslope scale [*McDonnell and Tanaka*, 2001; *Sivapalan*, 2003], there is an urgent need to fill knowledge gaps at the catchment scale [*Tetzlaff et al.*, 2008]. Furthermore, it is essential that scaling relations be developed so that the predictive capabilities of hydrologic models may accurately transcend scales [*Beven*, 2001].

So, what might be a way forward? One way may be to change the question that we are asking [*Sivapalan*, 2009] from "What ***is*** the response to land-use change?" to "What ***causes*** the response to land-use change?" While seemingly similar, this re-posing of the question in terms of the "causes" rather than "effects" takes the work in a fundamentally new direction—where catchment function, it's main stocks, flows and residence times of water become central to the determination of the hydrologic impacts of forest harvest (or some other land-use change). In this dissertation, we use stable isotopes of water (oxygen-18 and deuterium) to address fundamental questions of causation pertaining to forest harvest effects on hydrologic change. Our methodology integrates the classic paired-catchment approach with tracer techniques (combined with hydrometric analysis, hydrogeophysics [*Robinson et al.*, 2008], and plant ecophysiological monitoring [*Asbjornsen et al.*, 2011]) to shed light on subsurface flow pathways, storage volumes, and transit times for our research catchments in the Oregon Coast Range.

Our research catchments are located within the Alsea Watershed Study (AWS) in the Oregon Coast Range, USA. Conducted over a 13-yr period beginning in 1959, the AWS was one of the original paired-catchment studies created to assess the impacts of forest harvesting in the Western USA [*Stednick*, 2008]. Findings from the AWS spawned the first forest policies in the US created to protect water resources [*Stednick*, 2008]. The AWS was reinitiated in 2005 to test the effects of contemporary industrial forest harvesting practices using the paired-catchment framework [*Ice et al.*, 2007].

Our work also builds upon important work conducted 125 km south of my site in the Oregon Coast Range at Mettman Ridge near Coos Bay, Oregon where a solid conceptual model of runoff processes in steep, upland catchments (with similar geology, climate, and soils to the AWS) showed that groundwater played an active role in streamflow generation [*Anderson et al.*, 1997; *Montgomery et al.*, 1997]. This is something that has not been considered previously in paired-catchment studies in the Pacific Northwest (PNW). While we have much to build upon from the Coos Bay work, the Mettman Ridge study did not address vegetation and the process link between transpiration and subsurface storage. These relations are very poorly understood as revealed by recent work by *Brooks et al.* [2010]. Clearly, transpiration constitutes a significant portion of the water balance in forested catchments [*Calder*, 1998; *Vanclay*, 2009]. However, it is still unclear as to how vegetation water-use interacts with subsurface water stores and what implications this might have for streamflow generation. Recent efforts to better understand these interactions have focused on the links between transpiration and diurnal variations in streamflow during the growing season. In systems like the Alsea (where the seasonal climate results in a relatively rain-free growing season), the correlation between measured sapflow and diurnal streamflow is strong, but the strength of the correlation decreases through the summer [*Bond et al.*, 2002; *Wondzell et al.*, 2007]. More importantly, measurements by *Bond et al.* [2002] showed that trees within a narrow, near-stream “zone of influence” could account for the diurnal streamflow suppression. While the conceptual model *Bond et al.* [2002] used to explain these linkages places the near-stream vegetation within the alluvial floodplain (which provides an intuitive hydrologic connection to the stream via the surficial aquifer), work by *Barnard et al.* [2010] and *Wondzell et al.* [2007; 2010] provides evidence of other drivers of diel streamflow fluctuations that contests the near-stream “zone of influence” hypothesis. A critical next step for understanding the “how” of forest harvest response at this and other headwater sites in the PNW is to address how transpiration impacts catchment storage and baseflow generation based on catchment morphology and landscape position.

Linked to this coupling of vegetation and streamflow is the problem of scaling runoff processes, and the physical and biological controls on these processes, from hillslopes and small experimental catchments—where most process-based research efforts to date have been focused—to larger drainage basin scales where landscape management decisions are relevant [*Soulsby et al.*, 2006]. One promising new direction that began in Oregon’s Western Cascades [*McGuire et al.*, 2005] is the finding of scale-invariant relationships for streamwater mean transit time (MTT) that correlate physical attributes of the catchment (catchment geometry metrics and soil types) [*Broxton et al.*, 2009; *Capell et al.*, 2011; *Hrachowitz et al.*, 2009; *Rodgers et al.*, 2005; *Tetzlaff et al.*, 2009]. Mean transit time is a useful metric for developing such scaling relationships as it provides temporal characterization of catchment flowpath diversity and storage [*McDonnell et al.*, 2010]. Here, we intercompare MTT scaling relations of our research catchments in the Oregon Coast Range to those studied by *McGuire et al.* [2005] in the Western Cascades to identify first-order controls on catchment storage that may help in extrapolating forest harvest effects beyond the scale of our experimental catchments.

The central question for my dissertation is “Can we use a process-based approach to better understand ***how*** forest harvest impacts catchment function (flowpaths, storage, and residence times) at multiple scales?” Below we briefly describe the objectives and research hypotheses used in each chapter of this dissertation to tackle this central question, using **isotopes as the connective tissue between each of the project components.** To our knowledge, this is the first time such a methodology has been used to answer these questions in a forest harvesting context. It is hoped that by changing the question from “what” to “how”, we can move beyond the limitations of the paired-catchment study voiced by *Andreassian* [2004] and begin to advance our ability to predict response to forest harvest and other land-use changes.

## DESCRIPTION OF CHAPTERS

*Chapter 2: Similar catchment forms and rainfall-runoff responses hide radically different plumbing: (1) A top-down analysis of mean transit time ranges and scaling relations*

Chapter 2 uses a catchment intercomparison study to investigate the role of geology in controlling streamwater MTTs and their relationships with the surrounding landscape. We build on the work of *McGuire et al.* [2005], where they found that strong relations between flowpath length and flowpath gradient and MTT for a set of variably-sized (0.1 to 62 km2) headwater catchments in the Western Cascades Range of Oregon, USA. We compare their findings to our Alsea research catchments in the central Coast Range of Oregon—less than 140 km to the west.

*Chapter 3: Similar catchment forms and rainfall-runoff responses hide radically different plumbing: (2) A bottom-up catchment storage investigation*

Chapter 3 is a process-based assessment of the findings from the top-down intercomparison study reported in Chapter 2. We investigate how bedrock permeability influences catchment storage, MTTs, and MTT scaling relationships. We focus our work on the Alsea catchments in the Oregon Coast Range because the hydrological implications of the permeable bedrock geology in these catchments are still poorly-understood.

*Chapter 4: Where are diel streamflow fluctuations generated?: A test of the near-stream zone of influence hypothesis*

Chapter 4 reports the results of a novel manipulative experiment employed to test the "near-stream" zone of influence hypothesis poses by Bond et al. [2002]. We use isotope tracing of tree xylem water sources as an alternative, yet complementary means to test the hypothesis. We use our results explore the mechanisms responsible for generating diel streamflow fluctuations and discuss whether the further work in this area is warranted.

**REFERENCES**

Anderson, S. P., W. E. Dietrich, D. R. Montgomery, R. Torres, M. E. Conrad, and K. Loague (1997), Subsurface flow paths in a steep, unchanneled catchment., *Water Resour. Res.*, *33*(12), 2637-2653.

Andreassian, V. (2004), Waters and forests: from historical controversy to scientific debate, *Journal of Hydrology*, *291*(1-2), 1-27.

Asbjornsen, H., et al. (2011), Ecohydrological advances and applications in plant–water relations research: a review, *Journal of Plant Ecology*, *4*(1-2), 3-22.

Barnard, H. R., C. B. Graham, W. J. Van Verseveld, J. R. Brooks, B. J. Bond, and J. J. McDonnell (2010), Mechanistic assessment of hillslope transpiration controls of diel subsurface flow: a steady-state irrigation approach, *Ecohydrology*, *3*(2), 133-142.

Beschta, R. L. (1998), Forest hydrology in the Pacific Northwest: Additional research needs, *J. Am. Water Resour. Assoc.*, *34*(4), 729-742.

Beschta, R. L., M. R. Pyles, A. E. Skaugset, and C. G. Surfleet (2000), Peakflow responses to forest practices in the western cascades of Oregon, USA, *Journal of Hydrology*, *233*(1-4), 102-120.

Beven, K. J. (2001), How far can we go in distributed hydrological modelling?, *Hydrol. Earth Syst. Sci.*, *5*(1), 1-12.

Blöschl, G. (2001), Scaling in hydrology, *Hydrol. Processes*, *15*, 709-711.

Bond, B. J., J. A. Jones, G. Moore, N. Phillips, D. Post, and J. J. McDonnell (2002), The zone of vegetation influence on baseflow revealed by diel patterns of streamflow and vegetation water use in a headwater basin, *Hydrol. Processes*, *16*(8), 1671-1677.

Bosch, J. M., and J. D. Hewlett (1982), A REVIEW OF CATCHMENT EXPERIMENTS TO DETERMINE THE EFFECT OF VEGETATION CHANGES ON WATER YIELD AND EVAPO-TRANSPIRATION, *Journal of Hydrology*, *55*(1-4), 3-23.

Brooks, J. R., H. R. Barnard, R. Coulombe, and J. J. McDonnell (2010), Ecohydrologic separation of water between trees and streams in a Mediterranean climate, *Nature Geoscience*, *3*(2), 100-104.

Broxton, P. D., P. A. Troch, and S. W. Lyon (2009), On the role of aspect to quantify water transit times in small mountainous catchments, *Water Resour. Res.*, *45*.

Calder, I. R. (1998), Water use by forests, limits and controls, *Tree Physiology*, *18*(8-9), 625-631.

Capell, R., D. Tetzlaff, A. J. Hartley, and C. Soulsby (2011), Linking metrics of hydrological function and transit times to landscape controls in a heterogeneous mesoscale catchment, *Hydrol. Processes*, n/a-n/a.

Gomi, T., R. C. Sidle, and J. S. Richardson (2002), Understanding processes and downstream linkages of headwater systems, *Bioscience*, *52*(10), 905–916.

Hewlett, J. D., H. W. Lull, and K. G. Reinhart (1969), In Defense of Experimental Watersheds, *5*(1), 306-316.

Hibbert, A. R. (1967), Forest treatment effects in water yield, paper presented at Forest Hydrology, Pergamon Press, The Pennsylvania State University, University Park, PA, Aug. 29 to Sept. 10, 1965.

Hrachowitz, M., C. Soulsby, D. Tetzlaff, J. J. C. Dawson, and I. A. Malcolm (2009), Regionalization of transit time estimates in montane catchments by integrating landscape controls, *Water Resour. Res.*, *45*.

Ice, G. G., V. C. Hale, J. Light, J. J. McDonnell, S. H. Schoenholtz, and J. D. Stednick (2007), Testing contemporary forest practices: The Alsea Watershed Study Revisited., in *Sustaining America's Forests: Proceedings of the 2007 Society of American Foresters National Convention*, edited, Society of American Foresters, Portland, OR.

Jones, J. A., and G. E. Grant (1996), Peak flow responses to clear-cutting and roads in small and large basins, western Cascades, Oregon, *Water Resour. Res.*, *32*(4), 959-974.

Kovner, J. L., and T. C. Evans (1954), A Method for Determining the Minimum Duration of Watershed Experiments, *Transactions, American Geophysical Union*, *35*(4), 608-612.

McDonnell, J. J., and T. Tanaka (2001), Hydrology and biogeochemistry of forested catchments., in *Hydrol. Processes*, edited, John Wiley & Sons, New York.

McDonnell, J. J., et al. (2010), How old is streamwater? Open questions in catchment transit time conceptualization, modelling and analysis, *Hydrol. Processes*, *24*(12), 1745-1754.

McGuire, K. J., J. J. McDonnell, M. Weiler, C. Kendall, B. L. McGlynn, J. M. Welker, and J. Seibert (2005), The role of topography on catchment-scale water residence time, *Water Resour. Res.*, *41*(5), -.

Montgomery, D. R., W. E. Dietrich, R. Torres, S. P. Anderson, J. T. Heffner, and K. Loague (1997), Hydrologic response of a steep, unchanneled valley to natural and applied rainfall, *Water Resour. Res.*, *33*(1), 91-109.

Murtaugh, P. A. (2000), Paired intervention analysis in ecology, *Journal of Agricultural Biological and Environmental Statistics*, *5*(3), 280-292.

NRC (2008), *Hydrologic Effects of a Changing Forest Landscape*, National Acadamies Press, Washington, D.C.

Robinson, D. A., et al. (2008), Advancing process-based watershed hydrological research using near-surface geophysics: a vision for, and review of, electrical and magnetic geophysical methods, *Hydrol. Processes*, *22*(18), 3604-3635.

Rodgers, P., C. Soulsby, and D. Tetzlaff (2005), Using stable isotope tracers to assess hydrological flow paths, residence times and landscape influences in a nested mesoscale catchment, *Hydrol. Earth Syst. Sci.*, *9*, 139-155.

Seibert, J., and J. J. McDonnell (2010), Land-cover impacts on streamflow: a change-detection modelling approach that incorporates parameter uncertainty, *Hydrol. Sci. J.-J. Sci. Hydrol.*, *55*(3), 316-332.

Sidle, R. C. (2006), Field observations and process understanding in hydrology: essential components in scaling, *Hydrol. Processes*, *20*, 1439-1445.

Singh, V. P., and D. A. Woolhiser (2002), Mathematical modeling of watershed hydrology, *Journal of Hydrologic Engineering*, *7*(4), 270-292.

Sivapalan, M. (2003), Process complexity at hillslope scale, process simplicity at the watershed scale: is there a connection?, *Hydrol. Processes*, *17*(5), 1037-1041.

Sivapalan, M. (2009), The secret to 'doing better hydrological science': change the question!, *Hydrol. Processes*, *23*(9), 1391-1396.

Soulsby, C., D. Tetzlaff, S. M. Dunn, and S. Waldron (2006), Scaling up and out in runoff process understanding: insights from nested experimental catchment studies, *Hydrol. Processes*, *20*(11), 2461-2465.

Stednick, J. D. (1996), Monitoring the effects of timber harvest on annual water yield, *Journal of Hydrology (Amsterdam)*, *176*(1/4), 79-95.

Stednick, J. D. (Ed.) (2008), *Hydrological and Biological Responses to Forest Practices: The Alsea Watershed Study*, Springer, New York.

Tetzlaff, D., J. P. McNamara, and S. K. Carey (2011), Measurements and modelling of storage dynamics across scales, *Hydrol. Processes*, *25*(25), 3831-3835.

Tetzlaff, D., J. Seibert, K. J. McGuire, H. Laudon, D. A. Burns, S. M. Dunn, and C. Soulsby (2009), How does landscape structure influence catchment transit time across different geomorphic provinces?, *Hydrol. Processes*, *23*(6), 945-953.

Tetzlaff, D., J. J. McDonnell, S. Uhlenbrook, K. J. McGuire, P. W. Bogaart, F. Naef, A. J. Baird, S. M. Dunn, and C. Soulsby (2008), Conceptualizing catchment processes: simply too complex?, *Hydrol. Processes*, *22*(11), 1727-1730.

Thomas, R. B., and W. F. Megahan (1998), Peak flow responses to clear-cutting and roads in small and large basins, western Cascades, Oregon: a second opinion, *Water Resources Research*, *34*(12), 3393-3403.

Vanclay, J. K. (2009), Managing water use from forest plantations, *For. Ecol. Manage.*, *257*(2), 385-389.

Wondzell, S. M., M. N. Gooseff, and B. L. McGlynn (2007), Flow velocity and the hydrologic behavior of streams during baseflow, *Geophys. Res. Lett.*, *34*(24).

Wondzell, S. M., M. N. Gooseff, and B. L. McGlynn (2010), An analysis of alternative conceptual models relating hyporheic exchange flow to diel fluctuations in discharge during baseflow recession, *Hydrol. Processes*, *24*(6), 686-694.

Zegre, N., A. E. Skaugset, N. A. Som, J. J. McDonnell, and L. M. Ganio (2010), In lieu of the paired catchment approach: Hydrologic model change detection at the catchment scale, *Water Resour. Res.*, *46*.

# Similar catchment forms and rainfall-runoff responses hide radically different plumbing: (1) A top-down analysis of mean transit time ranges and scaling relations

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**INTRODUCTION**

Although it is widely-accepted that geology is a dominant control on catchment function [*Capell et al.*, 2011; *Jefferson et al.*, 2008; *Onda et al.*, 2006; *Tague and Grant*, 2004], the effect of geology and underlying boundary conditions on streamwater mean transit time (MTT) and MTT scaling relationships in upland catchments is relatively unknown. This knowledge gap exists despite growing bodies of ongoing work focused on two important research strands in watershed science: 1) the influence of catchment scale and structure on MTT [including *Hrachowitz et al.*, 2009; *McGlynn et al.*, 2003; *McGuire et al.*, 2005; *Rodgers et al.*, 2005; *Tetzlaff et al.*, 2009a] and 2) the role of geology and, in particular, bedrock groundwater in runoff generation processes [including *Anderson et al.*, 1997; *Asano et al.*, 2002; *Haria and Shand*, 2004; *Kosugi et al.*, 2006; *Kosugi et al.*, 2011; *Millares et al.*, 2009; *Soulsby et al.*, 2007; *Wilson and Dietrich*, 1987]. While these efforts have progressed rather separately, few studies yet have isolated bedrock geology to address its control on the partitioning, storage, and release of water scales from small to large catchments.

The need to develop scaling laws remains a grand challenge in hydrological research [*Blöschl*, 2001; *Dooge*, 1986; *Sivapalan*, 2003; *Tetzlaff et al.*, 2010]. Although scaling is an "umbrella" problem common to nearly all aspects of hydrology and related disciplines [*McGlynn et al.*, 2003], one of the principal foci is developing relationships that connect catchment function across multiple scales [*Tetzlaff et al.*, 2010]. Deriving scaling relationships that account for catchment function is a particularly important enterprise [*Sivapalan*, 2005] as it is a precursor for extending the process-based knowledge gained on experimental hillslopes and small catchments to larger, more management-relevant scales [*Tetzlaff et al.*, 2008]. Thus, the ability to develop predictive models that work for the "right reasons" [*Kirchner*, 2006] (and are applicable outside of the comforts of data-dense research catchments) hinges on advancing these functional scaling relationships beyond specific sites to more broadly generalizable attributes [*McDonnell et al.*, 2010; *Sivapalan*, 2003; *Soulsby et al.*, 2010]. Catchment geology may be such an attribute.

Geology's influence on physical hydrologic processes is multi-faceted (its role in biogeochemical processes is acknowledged but not covered here). As the parent material for pedogenic processes, geology influences soil hydraulic properties by setting the conditions that result in soil structure development and particle size distributions [*Jenny*, 1941; *Lohse and Dietrich*, 2005]. At the soil bedrock interface, bedrock type and weathering rate controls permeability contrasts which set lower boundary conditions within the subsurface [*Freeze and Cherry*, 1979]—dictating the length and conductivity of flowpaths and controlling the pore space available for water storage. These factors have been linked to stream MTT by several workers; *Asano et al.* [2002], and *Kabeya et al.* [2007] both showed that MTT was influenced by flowpaths through bedrock while *Uchida et al.* [2006] and *Katsuyama et al.* [2010] implicate bedrock permeability as a dominant control on MTT.

Stream water transit time is defined as the amount of time elapsed from the instant a molecule of water enters the catchment system as precipitation until it exits as streamflow (in most cases the "exit" point is defined by a sampling and/or gauging location) [*McGuire and McDonnell*, 2006]. Since a diversity of potential flowpaths exists within the catchment, an instantaneous parcel of stream water is properly defined by a distribution of travel times rather than a single value. The mean of this distribution (MTT) serves as a metric of catchment function as it integrates subsurface storage and transport properties into a single value [*Stewart and McDonnell*, 1991]. As a result of its inherent process-based representation at the catchment scale, MTT has become a primary hydrologic research tool used in studies aiming to better understand the controls on catchment function and their scaling relationships [*McDonnell et al.*, 2010; *Soulsby et al.*, 2009].

Initial investigations of MTT scaling behavior set out to test the intuitive hypothesis that MTT increases with catchment area [*McGlynn and McDonnell*, 2003; *McGuire et al.*, 2005; *Rodgers et al.*, 2005]. Although this hypothesis originated from preliminary findings by *Dewalle et al.* [1997] and *Soulsby et al.* [2000], to date, no studies have shown a strong correlation between catchment area and MTT [*Tetzlaff et al.*, 2009b]. Instead, researchers have found that MTT is more closely linked to landscape organization [*McGlynn et al.*, 2003], catchment topography [*McGuire et al.*, 2005], soil type [*Soulsby et al.*, 2006], or some combination thereof [*Hrachowitz et al.*, 2009; *Rodgers et al.*, 2005]. Although *Tetzlaff et al.* [2009b] compared relationships between MTT and landscape characteristics across five geomorphic provinces, to our knowledge, there have been no attempts to explore the explicit link between catchment geology and MTT across multiple scales.

Here we use a catchment intercomparison framework to isolate geology and subsurface boundary conditions to investigate their role in setting MTTs and their associated scaling relationships. We build on the work of *McGuire et al.* [2005], where they found that strong relations between flowpath length and flowpath gradient and MTT for a set of variably-sized (0.1 to 62 $km^2$) headwater catchments in the Western Cascades Range of Oregon, USA. We compare their findings to a neighboring set of research catchments in the central Coast Range of Oregon—less than 140 km to the west. As we will show, these catchments have nearly identical climate, topography, vegetation and physical hydrologic response to the *McGuire et al.* [2005] catchments. The one contrasting feature is the geology; permeable sedimentary geology in our Oregon Coast Range catchments compared to volcanic-dominated geology of the Oregon Cascades. We juxtapose these two catchments to test the null hypothesis that catchments having similar climate, form, vegetation, and rainfall-runoff response should have the same MTT ranges and scaling relationships—essentially asking the question “do differences in bedrock geology affect MTTs and their relationships with the overlying landscape?". Specifically, we address the following questions:

1) How do differences in geology (sandstone geology catchments of the Oregon Coast Range versus catchments of volcanic origin in the Western Cascades Range) compare in terms of physical morphometric properties?

2) How does geology affect hydrological flow regime?

3) How does geology affect streamwater MTT magnitude?

4) Do the relationships between stream MTTs and catchment landscape characteristics exhibit the same scaling behavior across these two contrasting geologies?

## STUDY SITES

*Drift Creek Catchments- Central Oregon Coast Range*

Our Coast Range study area is the upper Drift Creek basin (Figure 2.1; 44.5°N 123.9°W) within the Alsea River watershed in the central Oregon Coast Range. The Needle Branch catchment within the Drift Creek drainage has been gauged intermittently since 1959 as part of the Alsea Watershed Study (1959-1973) [*Harris*, 1977], and now part of the Alsea Watershed Study Revisited (2005-2019) [*Ice et al.*, 2007], to investigate the hydrological, biological, and water quality effects of forest management practices in the central Oregon Coast Range (http://www.ncasi.org/programs/areas/forestry/alsea/default.aspx).

Drift Creek is a $4^{th}$-order stream that flows in a southwesterly direction and joins the Alsea River near the head of its estuary approximately 5 km east of Waldport, Oregon (Figure 2.1). It drains a highly-dissected mountainous area, characterized by short, steep slopes that give rise to medium- to high-gradient streams

[*Thorson*, 2003]. Elevation within the Drift Creek catchment ranges from 110 to 857 m. Mean annual precipitation is 2500 mm based on the average of all cells of the PRISM 1971-2000 “normals” grid (PRISM Climate Group, Oregon State University, http://prism.oregonstate.edu, created 16 June 2006) contained within the Drift Creek watershed (800m cell size). On average, greater than 85% of the annual precipitation occurs from October through April in “long-duration, low-to-moderate intensity frontal storms” [*Harr*, 1976]. Snow accumulation occurs occasionally, but is typically highly transient. Studies in the past have neglected snow as part of the precipitation record [*Harris*, 1977].

The bedrock underlying the Drift Creek research area is the Eocene-aged Tyee Formation. The Tyee is composed of rhythmic-bedded layers of marine-derived greywacke sandstones and siltstones [*Snavely et al.*, 1964]. The beds range from 0.6 to 3.0 m and average 0.9 to 1.5 m thickness. A layer of saprolite, ranging from a few tenths to several meters thick (based on observations at well installations and road cuts), lies immediately below the soil profile. The shallow bedrock underneath the saprolite is highly fractured with fracture density decreasing with depth (Figure 2.2). This characteristic is corroborated by boring logs from a series of shallow (to 5 m) wells and one 35 m geotechnical hole at the nearby (120 km south of our site) Mettman Ridge research site near Coos Bay, Oregon [*Montgomery et al.*, 1997], which also overlies the Tyee Formation. *Snavely et al.* [1964] report that the porosity of the Tyee matrix ranges from 5 to 21% (mean=14%, n=17) and the permeability 2.2E-16 to 4.4E-15 $m^2$ (mean=2.7E-15 $m^2$). These permeability values are within the range of expected values for local permeability of fresh sandstone [*Freeze and Cherry*, 1979] and match recently reported landscape-scale estimates for this area [*Gleeson et al.*, 2011].

Upland soils within the Drift Creek research area are loams to gravelly loams (mesic Alic Hapludands and mesic Andic Humudepts) that average 1 m depth and are classified as well- to very well-drained [*Corliss*, 1973]. Field measurements of saturated hydraulic conductivity at the Needle Branch experimental catchment were not possible in most hillslope locations using a constant-head permeameter [*Amoozegar*, 1989] due to extremely high conductivities (that exceed field permeametry limits of ~1000 mm/hr). *Torres et al.* [1998] experienced the same problem using a Guelph Permeamter at Mettman Ridge. Hillslope soils at Mettman Ridge are the same mapped soil series as those in Drift Creek; therefore soil hydraulic properties measured at Mettman Ridge are assumed to be representative of those within Drift Creek. At Mettman Ridge, *Montgomery et al.* [1997] estimated average saturated conductivities of $10^{-3}$ m $s^{-1}$ in colluvial soil and $10^{-5}$ m $s^{-1}$ in the saprolite material forming the C-horizon using falling head tests in a series of piezometers (in=28 for soil and n=3 for saprolite). Also at Mettman Ridge, *Anderson et al.* [1997] reported porosities averaging 70% (n=12) and *Torres et al.* [1998] showed that these soils have very steep soil water retention curves—meaning that although they transmit water rapidly when at or near saturation, their hydraulic conductivity declines significantly with decreasing water potential and they retain little water relative to their total porosity. Valley bottom soils in the Drift Creek research area are silt loams (isomesic Fluvaquentic Humaquepts), that average 2 m depth and are classified as moderately permeable and somewhat poorly drained [*Corliss*, 1973].

Vegetation within Drift Creek is characterized by a patchwork of forest stands of varying maturity, linked to a mosaic of Federal and private industrial timberland ownership [*Stanfield et al.*, 2002]. The dominant canopy species is Douglas-fir (*Pseudotsuga menziesii*), but western hemlock (*Tsuga heterophylla*), western redcedar (*Thuja plicata*) and red alder (*Alnus rubra*) are present in varying degrees based on ownership and management intensity.

*HJ Andrews Experimental Forest Catchments- Central Oregon Cascades*

The HJ Andrews Experimental Forest (HJA; http://andrewsforest.oregonstate.edu/lter/) is located in the Western Cascades range near Blue River, Oregon, USA (Figure 2.1, 44.2°N 122.2°W). The HJA is generally defined by the Lookout Creek watershed boundary, although there are several small gauged catchments located outside, but adjacent to the Lookout Creek drainage area. Lookout Creek flows into the Blue River Reservoir just downstream of the HJA administrative boundary. Below the reservoir, the Blue River joins the McKenzie River which is a major drainage for the central Cascades Range.

Similar to the Coast Range catchments, the landscape of the HJA is steep, highly-dissected, and drained by medium- to high-gradient streams [*Thorson*, 2003]. Elevations within the study area range from 422 to 1628 m. Mean annual precipitation for the HJA is 2280 mm (also based on the average of all cells of the PRISM 1971-2000 "normals" grid contained within the area of interest). Precipitation timing is governed by the same Mediterranean-seasonality occurring in the Coast Range. However, the dominant precipitation phase varies more significantly in the Cascades as a result of the larger elevation gradient. Lower elevation catchments receive predominantly liquid precipitation, but also develop transient snow packs as approximately 25% of the total precipitation occurs as snowfall [*Bierlmaler and McKee*, 1989]. Higher elevation sites are snow-dominated and develop a seasonal snow pack, with peak snow water equivalent occurring from February to April [*Harr*, 1981; *Mazurkiewicz et al.*, 2008].

The bedrock geology consists of three mapped formations that occur as a function of elevation [*Swanson and James*, 1975]. At elevations less than 760 m, Oligocene to early-Miocene age hydrothermally-altered rock (massive breccias and tuffs) originating predominantly from mudflows and pyroclastic flows make up the

Little Butte Formation. At elevations ranging from 760 to 1200 m, Miocene ash flows and basalt and andesite lava flows comprise the Sardine Formation. The Pliocascade Formation, consisting of Pliocene to early-Miocene andesite lava flows, underlies elevations greater than 1200 m. The permeability of these volcanically-derived materials can be highly variable, but is generally a function of age [*Jefferson et al.*, 2010] and depth [*Saar and Manga*, 2004] as a result of hydrothermal alteration [*Ingebritsen et al.*, 1992]. Matrix porosity ranges from 2 to 10% and permeability ranges from 2.5E-14 to 5.0E-16 $m^2$ for the rock types and ages underlying HJA [*Ingebritsen et al.*, 1992]. These permeability values match well with reported values for local permeability and landscape-scale permeability estimated for this area [*Gleeson et al.*, 2011]. Fracturing associated with the cooling and shrinking of flow material is common near the top margin of individual flow units [*Peck et al.*, 1964], which can range from less than a meter to several tens of meters in thickness. Fractures connecting units vertically are associated with faulting [*Swanson and James*, 1975]. Boring logs from wells installed in a lower elevation catchment (Watershed 10, gauge elevation=462 m) and higher elevation catchment (Watershed 7, gauge elevation=938 m) both indicate high fracture densities near the bedrock surface that appear to decrease with depth (see Figure 2.2 for representative borehole diagrams from Watershed 10). Deep rock aquifers are present at the HJA (as observed from a drinking water well installed to 88 m), but little is known about the water source or flow directions within these deeper units.

Soils at the HJA vary based not only on landscape position, but also on landscape formation processes [*Swanson and James*, 1975]. In the steep, lower elevation catchments, loams to clay loams (Typic Dystrocryepts) of residual or colluvial origin and with average thickness of 1m or less are common [*McGuire and McDonnell*, 2010]. On topographic ridges, deep saprolite sequences can result in sub-soil depths of up to 7 m [*Harr*, 1977]. Despite the heterogeneity in profile thickness, hillslope soils at the HJA are highly permeable. In an intensive soil survey, *Ranken* [1974] measured saturated hydraulic conductivities in excess of $10^{-4}$ m $sec^{-1}$ for upper profile soils (A- and B-horizons, n=24 for each horizon). Mean porosity for these horizons was 65%, which decreases to 55% for the C-horizon. *Ranken* [1974] found a strong correlation between porosity and saturated hydraulic conductivity, meaning conductivity generally decreased with depth. Soil moisture exhibits a very steep non-linear relationship with matric potential in the near-saturation range, similar to the hillslope soils of the Drift Creek basin. Valley bottom soils are of alluvial origin, except in areas where landslide or debris flows have deposited colluvial material on top of the alluvial sediments, creating locally deep and variable soils [*Swanson and James*, 1975].

Vegetation cover varies by elevation at the HJA. At lower elevations, Douglas-fir, western hemlock, and western redcedar are the dominant canopy species. Noble fir (Abies procera), Pacific silver fir (Abies amabilis), Douglas-fir, and western hemlock dominate higher elevations.

## METHODS

We used a step-wise intercomparison approach to test the null hypothesis that catchments that have similar climate, form, vegetation, and rainfall-runoff response should have the same MTT ranges and scaling relationships. We use Needle Branch-12 (NB-12; number refers to catchment area in hectares) and Needle Branch-86 (NB-86) from the Drift Creek basin in the Coast Range and WS10 (10 ha) and WS01 (96 ha) from the HJA research area in the Western Cascades Range for direct comparison because of their similarities in size, topography, forest cover, and dominant precipitation type. We then compared the ranges of MTT and their scaling relationships from the larger nested study areas to see if differences in bedrock geology affect MTTs and their relationships with the overlying landscape.

*Terrain Analysis*

We used digital elevation models (DEM) with 10 m grid cells to derive indices of catchment form and organization. We used the D-infinity flow direction algorithm [*Tarboton*, 1997] to derive a flow accumulation grid. The area threshold method was used to define stream cells based on the flow accumulation, using an area threshold of 2.5 ha. This threshold value missed some 1$^{st}$-order channels as observed in the field, but was found to most closely match the actual stream network without creating a proliferation of “erroneous” channels (a problem encountered using smaller thresholds). The gridded stream networks were forced to agree with channels as mapped by the National Hydrography Dataset stream lines (http://nhd.usgs.gov), verified in the field with a global positioning system, for Drift Creek and a channel map derived from airborne laser altimetry points and field mapping for HJA (personal communication Theresa Valentine).

We used the DEM, flow accumulation grid, and delineated stream network to calculate slope ($S$), drainage density ($D_d$), and area-to-perimeter ratio ($A$-$P$) for each study catchment. Additionally, we calculated subcatchment area ($SCA$) [*McGlynn and Seibert*, 2003], defined as the median value of the accumulated area draining to each individual stream cell within a catchment. The D-infinity algorithm splits flow between cells resulting in multiple possible flow paths leading to a single cell. We therefore used a weighted average of flow path lengths to each cell to calculate flowpath length ($L$) and flowpath gradient ($G$). We also calculated the topographic wetness index ($TWI$) [*Beven and Kirkby*, 1979b],

$$TWI = \ln\left(\frac{a}{\tan\beta}\right) \quad (1)$$

where $a$ is the upslope contributing area per contour interval and $\tan\beta$ is the local slope, and the downslope index [*Hjerdt et al.*, 2004]. The downslope index describes the horizontal distance, $L_d$, necessary to drop $d$ elevation units along the path of steepest descent in a catchment and is reported as the gradient:

$$DSI_d = \frac{d}{L_d}. \quad (2)$$

We used $d$=5 m as that value was determined to be indicative for steep terrain [*Hjerdt et al.*, 2004] and to remain consistent with other intercomparison studies [*Tetzlaff et al.*, 2010].

*Hydrometric Analysis*

We performed hydrometric analyses using hourly and daily streamflow and precipitation ($Q_h$, $Q_d$, $P_h$, and $P_d$, respectively) records from October 1, 2005 through September 30, 2009, with the exception of NB-12 where gauging did not begin until October 1, 2007. Precipitation inputs for NB-12 and NB-86 were taken as the areal average of a spatially distributed network of rain gauges present near the NB catchment (shown in Figure 2.1). The precipitation input for WS10 and WS01 was measured at the PRIMET meteorological station, which is the closest measurement point at the HJA (horizontally and in elevation; also shown in Figure 2.1).

Many statistics have been developed within the field of hydrology and its related disciplines to characterize streamflow regimes [*Olden and Poff*, 2003; *Wagener et al.*, 2007]. For our intercomparison, we used basic hydrological statistics as well as several indices of hydrodynamic response that were both relevant to our objectives and applicable based on the length of our data record. Basic statistics, such as mean annual flow (*MAF*), mean annual peak flow (*MAPF*), mean annual low flow (*MALF*), and coefficient of variation of stream discharge ($CV_Q$), were computed for each stream using the daily discharge record. Mean runoff ratio ($R_{QP}$) is the ratio of total discharge to total precipitation, calculated as,

$$R_{QP} = \frac{\sum Q_d}{\sum P_d} \tag{3}$$

and is a measure of the amount of incident precipitation that leaves the catchment as streamflow over the period of interest [*Sawicz et al.*, 2011].

We used the baseflow index (*BFI*) as a metric to relate the amount of discharge occurring as baseflow to the total amount of discharge over the period of interest [*Arnold et al.*, 1995]. Although many baseflow separation techniques exist, we used the constant-slope method of *Hewlett and Hibbert* [1967] to be consistent with other forested headwater catchment research whereby a line with slope 0.55 L $s^{-1}$ $km^{-2}$ $hr^{-1}$, beginning at the first point of storm response, separates event runoff ($Q_{evt}$) from baseflow ($Q_{bf}$). After $Q_{bf}$ separation, BFI was calculated as,

$$BFI = \frac{\sum Q_{bf}}{\sum Q_h}. \tag{4}$$

We used flow duration analysis to calculate the percent of time that stream discharge of a given magnitude will be equaled or exceeded [*Dingman*, 2002]. The local slope of the resulting flow duration curve (*FDC*) can be used as a measure of the degree of discharge variability for different magnitudes of the flow regime [*Sawicz et al.*, 2011]. Therefore, an index of discharge variability for intermediate flow ranges is the slope of the FDC between the 33$^{rd}$ and 66$^{th}$ flow percentiles and is calculated as,

$$FDC_{33-66} = \frac{\ln(Q_{33}) - \ln(Q_{66})}{0.66 - 0.33}. \tag{5}$$

We compared $Q_{evt}$ to total precipitation for a given event ($P_{evt}$) as a way to directly assess catchment response to precipitation inputs [*Graham and McDonnell*, 2010]. Our event rule specified that 5 mm or more of precipitation during a twelve-hour period was required to initiate the delineation of a precipitation event. Events were considered separate when a period of at least ten hours with mean precipitation intensity less than 0.1 mm $hr^{-1}$ occurred between them. We calculated event discharge as the sum of $Q_{evt}$ associated with each delineated precipitation event (separation was carried out using the previously described constant-slope method). We also used the Richards-Baker flashiness index [*Baker et al.*, 2004] as a metric of catchment responsiveness. The index, $FI_{RB}$, is calculated as,

$$FI_{RB} = \frac{\sum_{i=1}^{n} |Q_i - Q_{i-1}|}{\sum_{i=1}^{n} Q_i} \tag{6}$$

where $Q_i$ is the hourly discharge ($Q_h$) at each timestep, *i*.

We used streamflow recession analysis as a way to characterize how a catchment releases stored water [*Brutsaert and Nieber*, 1977]. By plotting the slope of the recession, $-dQ_d/dt$, versus $Q_d$ in log-log space, time dependence was removed from the analysis, therefore,

$$-\frac{dQ}{dt} = f(Q) \tag{7}$$

where *f* can be described as a power-law function [*Rupp and Woods*, 2008],

$$f(Q) = aQ^b\ . \tag{8}$$

The exponents, *b*, of the resulting fits were used to compare the behavior of the $-dQ/dt$-$Q$ relationships.

*Transit Time Estimation*

Mean transit times of stream baseflow were estimated by *McGuire et al.* [2005] for seven catchments within the HJA research area (WS02, WS03, WS08, WS09, WS10, MACK, and LOOK) using water isotopes in combination with lumped-parameter convolution models following the methodology of *Maloszewski and Zuber* [1982]. Water isotopes, oxygen-18, deuterium, and tritium, are ideal hydrological tracers as they are a part of the water molecule (rather than a separate molecule as typical of most artificial tracers) and, consequently, are fully conservative. Estimating MTT using the convolution approach assumes that the isotopic composition of the water coming out of the system, $\delta_{out}$ (stream water), will be equal to the composition of the water coming in, $\delta_{in}$ (precipitation), lagged by some time, τ, and weighted by the transit time distribution, $g(\tau)$, and recharge weighting function $w(t-\tau)$. This is expressed mathematically as,

$$\delta_{out}(t) = \frac{\int_0^\infty g(\tau)w(t-\tau)\,\delta_{in}(t-\tau)d\tau}{\int_0^\infty g(\tau)\,w(t-\tau)d\tau}. \tag{9}$$

The $g(\tau)$ term describes the density and range of transit times within the catchment [*McGuire and McDonnell*, 2006] while the $w(t-\tau)$ function is used to conserve tracer mass in the system by weighting $\delta_{in}$ according to the fraction of precipitation estimated to be contributing to recharge [*Stewart and McDonnell*, 1991].

We used the same approach of *McGuire et al.* [2005] to estimate MTT for eight nested catchments in the Drift Creek basin (Figure 2.1; NB-12, NB-34, NB-86, FC-210, DC-315, MC-1881, DR-5373, and DR-8643). We used the deuterium ($\delta^2$H) composition of precipitation and stream waters as $\delta_{in}$ and $\delta_{out}$, respectively, in the lumped-parameter convolution models. We collected bulk precipitation samples at locations representing low-, mid-, and high-elevations within the Drift Creek basin on weekly to b-weekly intervals from January 2006 through September 2010. Stream samples were collected at each sampling location on near-weekly intervals beginning in July 2007. Stream samples collected during stormflow conditions were not included in our analysis to remain consistent with the methodology of *McGuire et al.* [2005].

Water samples were analyzed for $\delta^2$H composition using off-axis integrated cavity output laser spectroscopy on a Los Gatos Research Liquid Water Isotope Analyzer (LWIA-24d, Los Gatos Research, Inc.) at the Oregon State University Water Isotope Collaboratory. Deuterium values were reported as ratios relative to Vienna Standard Mean Ocean Water (V-SMOW) in standard "delta" notation,

$$\delta^2\mathrm{H}\ [‰] = (\frac{R_{sample}}{R_{standard}} - 1) \times 1000 \tag{10}$$

where $R_{sample}$ and $R_{standard}$ are the ratios of $^2$H/$^1$H of the sample and V-SMOW, respectively. Lab standards were routinely verified using isotope ratio mass spectrometry (multiple labs). Analytical precision for $\delta^2$H was 1.0‰.

Mean transit time estimation was accomplished using an inverse modeling procedure where the modeled $\delta^2$H composition of streamflow was fitted to the measured $\delta^2$H composition by iteratively adjusting the parameters of the transit time distribution, $g(\tau)$. We used the gamma model to approximate $g(\tau)$ as it has been shown to be more theoretically representative of real catchment systems than the more commonly used single-parameter exponential model [*Kirchner et al.*, 2000]. Due to the flexibility gained with the addition of a second parameter, the gamma distribution can better handle a diversity of both short and long flowpaths. It is modeled as,

$$g(\tau) = \frac{\tau^{\alpha-1}}{\beta^{\alpha}\Gamma(\alpha)} exp^{-\frac{\tau}{\beta}} \tag{11}$$

where α is the shape parameter, β is the scale parameter, and the MTT is equal to αβ.

Parameter and predictive uncertainty was estimated using a Bayesian approach [e.g. *Hrachowitz et al.*, 2010] where the posterior distribution, $p(\theta, \sigma|Y)$, of the parameters, θ, are related to the likelihood function, $p(Y|\theta,\sigma)$, and error model, $p(\theta,\sigma)$, as,

$$p(\theta,\sigma|Y) \propto p(Y|\theta,\sigma)p(\theta,\sigma) \tag{12}$$

and σ is the standard deviation of the error model. The prior distributions of the parameters were defined as 0<α<3 and 0<β<30000 and assumed to be distributed uniformly. The likelihood function is expressed as,

$$p(Y|\theta,\sigma) = \prod_{i=1}^{N_y} N(\xi_i(\theta,Y)|0,\sigma^2) \tag{13}$$

where $N(\xi_i(\theta,Y))$ is the distribution of the residual errors, ξ, assumed independent and following a normal distribution with mean, μ, and variance, $\sigma^2$.

We used a Markov Chain Monte Carlo (MCMC) search procedure implemented within the DREAM-ZS algorithm [*Schoups and Vrugt*, 2010] to sample the prior parameter distribution. We ran the search procedure for 15000 iterations using 3 parallel chains to find the parameter set that maximized the log-likelihood function. Results from the first 5000 iterations were discarded as these were considered a warm-up period for the search algorithm. The remaining 10000 iterations were used in the uncertainty estimation.

## RESULTS

### *Catchment Form*

Catchment form was very similar for the Drift Creek and HJA catchments. Catchment area ($A_c$), minimum elevation ($E_{min}$), elevation range ($E_{rng}$), $D_d$, and *A-P,* as well as the median values of *S*, *L*, *G*, *SCA*, *TWI*, and $DSI_5$, for NB-12, NB-86, WS10, and WS01 are presented in Table 1. These values show that the research catchments were steep (median S ranges from 34 to 68 %) and highly dissected ($D_d$ ranges from 2.53 to 3.72 km-1). They also show that Drift Creek and HJA share similar organizational patterns when grouped by size, as measured by the *A-P* ratio and median *SCA* metric. The *A-P* ratio for the small catchments varied by only 0.01 km and the median *SCA* ranged from 4.1 ha at NB-12 to 7.2 ha at WS10. The larger catchment *A-P* ratios were 0.15 and 0.175 km for NB-86 and WS01, respectively, while their median *SCA* differed by only 2 ha (12.3 and 10.3 ha, respectively).

Further demonstration of the similarities in catchment form is provided in the calculated cumulative distribution functions (CDF) of *S*, *L*, *G*, *SCA*, *TWI*, and $DSI_5$ (Figures 2.3 (a-f). Although the shapes of the CDFs for *S* did not conform to one another exactly (Figure 2.3a), the CDFs for the *G* metric (Figure 2.3b), which is defined as the slope measured along the topographic flowpath, shared the same steep characteristic behavior (inter-quartile ranges (IQR) are 0.12, 0.17, 0.17, and 0.15 for NB-12, NB-86, WS10, and WS01, respectively). Likewise, the shape of the CDF of *L* was similar for all catchments, whereby *L* was accumulated in the same linear fashion to a density of approximately 0.80, at which point the rate of accumulation decreased significantly (Figure 2.3c). The shape of the CDF for *SCA* was dependent on catchment area; the smaller catchments followed a steep linear profile (NB-12 and WS10) and the larger catchments CDF shape took a more logarithmic form (Figure 2.3d). The *TWI* and $DSI_5$ are slope-dependent metrics, so it is no surprise that their CDFs also shared the same general shape across each of the catchments (Figure 2.3e and f).

*Hydrodynamic Response*

Daily precipitation and mean daily discharge for NB-12, NB-86, WS10, and WS01 are shown in Figure 2.4 (a and b, respectively). In general, the hydrographs displayed the same seasonality with distinct high flow and low flow periods. The high flow periods were marked by rapid rainfall-runoff response and the low flow periods were characterized by a gradual decline to annual minima. Precipitation totals were strikingly similar for the Oregon Coast Range and Western Cascades catchments in water years 2006 and 2007 (difference of 122 and -43 mm, respectively, Coast Range minus Cascades), but the Oregon Coast Range was drier in water years 2008 and 2009 (by 569 and 542 mm, respectively) (Figure 2.5a). Discharge totals for each year of record are shown in Figure 2.5b.

Table 1 presents the hydrologic statistics and hydrodynamic response metrics calculated for each catchment. The average volume of water discharged over the period of record for each research catchment varied by only 0.75 mm $d^{-1}$, as indicated by the *MAF* statistic. Peak discharges were also fairly equivalent during the study period with a high *MAPF* of 65.4 mm $d^{-1}$ at NB-12 and low *MAPF* of 60.5 mm $d^{-1}$ at WS01. Minimum discharges were similar for NB-86, WS10, and WS01 (*MALF*=0.04, 0.02, and 0.03 mm $d^{-1}$, respectively); *MALF* at NB-12 was substantially higher relative to the other catchments (0.34 mm $d^{-1}$). The $CV_Q$ ranged by 0.24 (from 1.76 at NB-86 to 1.99 at WS10), which is a direct reflection of the similarity in seasonal flow regime (as $CV_Q$ is a measure of the variation around the mean). The $R_{QP}$ values (Table 1; ranging from 0.59 to 0.95) indicated that, over the period of record, the majority of rainfall was converted to runoff for all catchments. The fraction of runoff that occurred as baseflow, estimated as the *BFI*, ranged from 0.62 (WS10) to 0.72 (NB-86), a difference of only 0.10. Hydrograph flashiness, as quantified by $FI_{RB}$, was also very similar amongst the catchments, with values ranging from 0.33 (NB-86) to 0.45 (WS10).

Flow duration analysis provided further indication of the similarity in flow regimes between the sedimentary and volcanic catchments. The flow duration curves showed that the distribution of discharge magnitudes were comparable across catchments for the ranges of flows measured during the study period (Figure 2.6). Additionally, the rainfall-runoff dynamics, as represented by the relationship between $Q_{evt}$ and $P_{evt}$, indicated that the catchments responded similarly to precipitation inputs (Figure 2.7). The baseflow recession behavior, captured by the *dQ/dt-Q* relationship, exhibited the same general pattern across each of the research catchments (Figure 2.8).

*Mean Transit Time Ranges and Landscape Relationships*

Precipitation $\delta^2$H values measured within the Drift Creek basin ranged from -120 to -10‰ and varied seasonally (lower values typically occurred during the colder winter months). No elevation effect was observed ($r^2$=0.03), so values from each bulk precipitation collector were used to create a spatially-averaged precipitation $\delta^2$H record as the model input, $\delta_{in}$. Stream $\delta^2$H values ranged from -55 to -45‰ and were highly-damped compared to the precipitation record.

A model warm-up period of 15 years was used to "prime" the model before fitting the measured $\delta_{out}$ record [following *Hrachowitz et al.*, 2010]. The warm-up dataset was created by first extending the $\delta_{in}$ to the beginning of the 2006 water year using regression relationships ($r^2$=0.63) between $\delta_{in}$ and $\delta^2$H values of precipitation collected at the US Environmental Protection Agency office in Corvallis, OR (Renee Brooks, unpublished data). The $\delta_{in}$ record was then looped three times to create the 15-yr dataset and appended to the beginning of the calibration dataset. We report MTT as the product of the parameter set, αβ, having the maximum log-likelihood value ($MTT_{mle}$, Table 2).

We compared our MTT estimates from the Drift Creek basin (Table 2) to values reported by McGuire et al. (2005) for seven catchments within the HJA (Table 3). MTTs were longer in the sedimentary Coast Range catchments (3.7 to 10.4 yrs) than they were in the volcanic Western Cascades catchments (0.8 to 3.3 yrs). Excluding WS08 (MTT=3.3 yrs), which has deeper soils (>3 m) and is lower gradient (mean slope=30%) than the other HJA catchments *McGuire et al.* [2005] characterized, MTTs ranged from only 0.8 to 2.2 yrs. Following our direct comparison from the previous two sections, the MTTs for NB-12 and NB-86 were 5.0 and 4.0 yrs; distinctly longer than that reported for WS10 (1.2 yrs; Table 2) and WS01 (1.3 yrs[1]).

Unlike the *McGuire et al.* [2005] findings at the HJA, MTTs for the sedimentary Drift Creek catchments were not significantly correlated to median slope length $L$, median slope gradient $G$, or $L/G$ at an alpha level of 0.05 (Figure 2.9). Instead, we found a significant positive relationship between MTT and basin area, $\log(A_c)$ ($r^2$=0.67, p=0.01), at Drift Creek. The logarithm of $A_c$ ($\log(A_c)$) was used in the analysis (rather than $A_c$) to better meet the normality assumption of the regression model, although both showed the same positive relationship. Pearson's correlation [1]coefficients, r, and associated p-values for each MTT-landscape metric relationship are provided in Table 4. In addition to $A_c$, MTT was positively correlated to $A$-$P$ (r=0.91, p<0.01), a metric of catchment shape, at Drift Creek. At the HJA, median $S$ was negatively correlated to MTT (r=-0.86, p=0.03), which is not surprising given the topographic dependence already reported by *McGuire et al.* [2005]. Likewise, $TWI$ and $DSI_5$, also metrics of catchment topography, showed positive correlations (r=0.65, p=0.16 for each). Drainage density ($D_d$) was positively correlated to MTT at the volcanic HJA (r=0.69, p=0.13).

## DISCUSSION

### *Similar forms and response do hide radically different plumbing*

Our results clearly show that despite striking similarities in catchment form and hydrologic regime, MTT ranges and scaling relationships differ between the sedimentary Coast Range (Drift Creek) and volcanic Western Cascades (HJA)

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[1]McGuire et al. [2005] did not estimate MTT for WS01. We therefore used the regression relationship between MTT and the ratio of median flowpath length, L, and median flowpath gradient, G, to estimate an MTT of 1.3 yrs for WS01 (MTT=0.0021*(L/G)+0.71; $r^2$=0.91).

catchments. Mean transit times are up to an order of magnitude longer in the sedimentary catchments and increase with increasing basin area, whereas the relatively shorter MTTs of the volcanic catchments are strongly dependent on catchment topography and unrelated to basin area. The longer MTTs of the sedimentary catchments indicate that slow, and presumably deep, flowpaths play an important role in supplying water for streamflow. Further, the positive relationship between MTT and catchment area (Table 4) suggests the presence of flowpaths that not only age in a down-valley direction, but also contribute proportionately more water to the stream in a down-valley direction. This contrasts with the effect of the tight volcanic bedrock of the Western Cascades which induces shallow lateral subsurface flow [*Harr*, 1977; *McGuire and McDonnell*, 2010]. This runoff generation mechanism matches well with the strong correlation found between MTT and L/G at HJA, as transport via lateral subsurface flow is dependent on both L and G. These findings illustrate that similar catchment forms and rainfall-runoff responses can hide radically different subsurface plumbing, both in terms of absolute water age and how such measures scale across the landscape.

*Geology and physical morphometric properties in context*

Notwithstanding the distinct differences in geological composition, as well as landscape age and formation processes, the physical form of the Drift Creek basin in the Oregon Coast Range is remarkably similar to that of the HJA research area in the Western Cascades. In both research areas, the highly-dissected landscape is a result of fluvial incision and colluvial processes—namely shallow landsliding [*Dietrich and Dunne*, 1978; *Jefferson et al.*, 2010]. Although the landscapes at each study site are dominated by short, steep hillslopes, lower gradient slopes do exist and, in both cases, are primarily attributable to deep-seated landslides [*Roering et al.*, 2005; *Swanson and James*, 1975]. Lower gradient terrain along the periphery of the lower catchment boundary at NB-86 is associated with ancient deep-seated landslides and accounts for the lower median *S* value; this morphology is not present to such a significant extent in NB-12, WS10, or WS01. At the HJA, deep-seated landslides are more common in higher elevation catchments such as WS08 [*Swanson and James*, 1975].

These common erosional pathways have led to the development of dense drainage networks at each research site ($D_d$ is greater than 2.5 km-1 at both sites). *Jefferson et al.* [2010] used a chronosequence of catchment ages across the older Western Cascades and the younger High Cascades to show that landscape-scale drainage density is dependent on the age of the geological substrate. Shallow runoff processes become more dominant in the volcanic rocks of the Cascades as hydrothermal-alteration proceeds to make the rocks less permeable with increasing age, which, in turn, leads to enhanced fluvial-colluvial sculpting of the landscape. These findings illustrate the intimate link between geological and hydrological factors in landscape evolution. We therefore hypothesize that the differences in geologic substrate between the Oregon Coast Range and Western Cascades are trumped by physical weathering processes and have resulted in landscapes with common form.

*An ex post facto analysis of flow regime*

Considering the similar climate, vegetation, and topography of the Drift Creek and HJA catchments, it is perhaps not surprising that these catchments exhibited such similar hydrological regimes. However, given such large differences in MTTs, we were surprised that no obvious clues to these MTT differences were apparent in the analysis. Here we present a brief ex post facto examination of the hydrological behavior of NB-12, NB-86, WS10, and WS01 to see if any subtle differences can be identified and possibly attributed to the contrasting geologies. Although the discharge traces are indeed outwardly quite similar, two distinctions might be made. The most obvious difference is that summer baseflow in NB-12 remains elevated relative to its downstream counterpart (NB-86), WS10, and WS01 (Figure 2.4). This is reflected in the high $R_{QP}$ at NB-12 (0.95). To put this $R_{QP}$ value into perspective, *Sawicz et al.* [2011] showed that only a small fraction (<3%) of the 280 eastern United States catchments they surveyed had an $R_{QP}$ greater than 90%. Elevated summer baseflow paired with such an extreme $R_{QP}$ value may signify interbasin groundwater contributions. This scenario is plausible based on the tilted, layered geology with fracturing known to occur along the bedding planes in the Tyee Formation of the Oregon Coast Range [*Snavely et al.*, 1964]. The second difference is that the troughs of the WS10 and WS01 hydrographs seem to be deeper during the inter-storm period, potentially indicating, continued contributions via slower flowpaths after the initial runoff response in the sedimentary Coast Range catchments (NB-12 and NB-86). While this behavior is consistent with the longer MTTs estimated at these sites, a more thorough hydrometric investigation—including monitoring of the various groundwater stores contributing to runoff—is needed to fully characterize how the contrasting geologies (and MTTs) can lead to such similar hydrological flow regimes.

*The geology of MTT scaling*

Although geology has been implicated as a control on MTT in previous studies [*Katsuyama et al.*, 2010; *Uchida et al.*, 2006], no other study to date, that we are aware of, has shown such stark differences in MTTs and with nearly-opposite scaling behavior as we have observed with our sedimentary Oregon Coast Range and volcanic Western Cascades catchments. In a nested-catchment study within the River Feugh basin in northeast Scotland, *Rodgers et al.* [2005] found that soil type and topography were the best predictors of MTT; catchments dominated by peat soils had shorter MTTs than steeper catchments having more freely-draining soils and significant valley-bottom groundwater storage. Similarly, *Tetzlaff et al.* [2009a] found that soil type and topography were the primary controls on MTT in three sets of nested catchments in the Cairngorm Mountains of Scotland, where spatially-variable glacial drift deposits influence soil responsiveness. *Hrachowitz et al.* [2009] extended the

*Rodgers et al.* [2005] and *Tetzlaff et al.* [2009a] analyses across seven geomorphologically and climatologically distinct zones in the Scottish Highlands. Using multiple regression, *Hrachowitz et al.* [2009] showed that the extent of responsive soil cover, drainage density, precipitation intensity, and topographic wetness index explained 88 % of MTT variability in 20 catchments of varying size (<1 to 35 $km^2$). Contrasting with these examples, catchment organization (measured as the median sub-catchment area) was found to be the dominant control on MTT at the continuously wet Maimai catchments on the South Island of New Zealand [*McGlynn et al.*, 2003]. There, they found that MTT increased as sub-catchment area increased but that MTT and total catchment area were not correlated. Despite the potpourri of MTT scaling studies now in the literature, Drift Creek is the first basin (with an adequate number of nested catchments) to exhibit the area-dependence for MTT that motivated the first MTT scaling studies [*McGlynn and McDonnell*, 2003; *McGuire et al.*, 2005; *Rodgers et al.*, 2005].

Our study design is unique in that the Drift Creek and HJA catchments were so physically similar that we were able to explicitly “control” for the effect of bedrock geology—a rare occurrence in catchment studies. Therefore, it is easy to infer that the disparities in MTTs and their scaling relationships with landscape variables are a result of the differences in bedrock geology between the Oregon Coast Range and Western Cascades catchments; but the obvious next question is how, mechanistically, geology controls the subsurface storage, flowpaths and flow source of water? In landscapes with steep terrain, catchment flowpath distributions are expected to be a function of topography [*Beven and Kirkby*, 1979a; *McGuire et al.*, 2005]. While the scaling relations at HJA followed this expected trend, the Drift Creek sites in the sedimentary Oregon Coast range did not. Perhaps the biggest difference then between the two sites is their level of landscape-scale anisotropy. Anisotropy is a term used to quantify the relative heterogeneity, in the vertical and horizontal directions, of hydraulic conductivities in porous media [*Marcus and Evenson*, 1961]. This new notion of landscape-scale anisotropy is supported by data presented in *Uchida et al.* [2006] where bedrock permeability played an important role in controlling MTTs at the hillslope scale based on an intercomparison of the Maimai (New Zealand) and Fudoji (Japan) catchments. Similarly, *Katsuyama et al.* [2010] showed that bedrock permeability was a key factor in setting the streamwater MTTs in the granite-dominated bedrock catchments of the Kiryu Experimental Watershed in Japan. Theoretically, bedrock permeability should determine the presence (and density) of flowpaths below the soil profile while the sharpness of the permeability contrast between soil and rock should control the degree of partitioning between shallow and deep water flow paths. Since the matrix permeability of the Tyee sandstone at Drift Creek is similar to that of the various volcanic rock types at the HJA, we hypothesize that differences in effective bedrock permeability at the landscape scale, as well as the sharpness of permeability contrasts are both a function of the degree of weathering and fracturing within the bedrock. Bedrock weathering and bedrock fractures therefore appear to be a dominant control on streamwater MTTs and how they scale with the surrounding landscape for our two catchments. This processes linking bedrock permeability and MTTs are further explored in *Hale et al.* [this issue].

*On the assumptions and limitations of our approach*

The application of stable isotopes in lumped-parameter convolution models have theoretical limitations based on the transit time distributions selected [*Stewart et al.*, 2010]. While the gamma model allows characterization of slower flowpaths via its long-tail (relative to other distributions), our dataset pushes the limit of its application. In addition, the simplification of the catchment system as assumed in the lumped-parameter convolution models inherently introduces additional uncertainty into the MTT estimations (see *McGuire and McDonnell* [2006] for a detailed assessment of the assumptions). It is therefore appropriate to consider our MTT estimations indicative rather than absolute, as voiced by *Soulsby et al.* [2010]. Keeping this in mind, there is a significantly large degree of separation between the Drift Creek and HJA MTTs so that we stand confidently by our finding of longer MTTs in the sedimentary catchments relative to that of the volcanic catchments. We confirm this using a two-sample, single-tailed t-test to show that, indeed, the mean of the lower MTT uncertainty bounds at Drift Creek, 5.0 yrs, is significantly larger than the mean of the upper MTT uncertainty bounds at HJA, 2.4 yrs (p-value=0.005, t-stat=-2.95, 13 degrees of freedom). The robustness of the positive MTT-catchment area relationship, given the non-trivial MTT uncertainty at the Drift Creek catchments, was checked by conducting 1000 regressions where, in each case, the MTT values were randomly sampled from their uncertainty range. Of all MTT combinations, 78 % resulted in positive MTT-catchment area relationships significant at the 0.05 alpha level. Significant ($p<0.05$) regression coefficients ranged from 1.3 to 5.0 and non-significant ($p>0.05$) regression coefficients ranged from 0.5 to 3.4, both providing additional confidence in our finding of the positive MTT-catchment area relationship in the sedimentary Drift Creek basin.

Despite the strong inference of bedrock permeability as the primary control on MTT and MTT scaling relationships for these two research catchments, our top-down approach does not allow for a process-based explanation of 1) how bedrock geology influences MTT or 2) how catchments with such starkly different MTTs could have such similar hydrologic flow regimes. Therefore, it is still unclear mechanistically how the sedimentary catchments of the Oregon Coast Range have longer MTTs and why they scale differently than those of the volcanic Western Cascades at the HJA. *Hale et al.* [this issue] address these questions using a detailed field-based study to better understand the role of subsurface catchment storage on setting streamwater MTTs in the Oregon Coast Range.

**CONCLUSIONS**

We used a catchment intercomparison study to investigate the role of geology in controlling streamwater MTTs and their relationships with the surrounding landscape for a catchment with sedimentary bedrock in the Oregon Coast Range and a catchment with volcanic bedrock geology in the nearby Western Cascades. We showed that MTTs were longer in the Coast Range catchments with more permeable fractured and weathered sandstone bedrock than in the Western Cascades catchments with tight, volcanic bedrock. In the permeable bedrock catchments, MTT was positively correlated to basin area, whereas MTT was most strongly linked to the ratio of median flowpath length to median flowpath gradient in the volcanic catchments. Despite the differences in MTT magnitude and scaling relationships, the catchments displayed remarkable similarities in landscape morphometry and hydrological flow regimes. We therefore conclude that similar catchment forms and hydrologic responses can indeed lead to different MTTs and MTT scaling relationships.

Increasingly, regionalization is being used as an approach to parameterize models when calibration data is not available [*Yadav et al.*, 2007]. Topographic metrics and hydrodynamic response indices are the primary variables used in building the regression models that predict parameter sets across a region. Although many indices currently used in regionalization studies are considered to capture catchment function [*Oudin et al.*, 2010; *Sawicz et al.*, 2011; *Yadav et al.*, 2007], our results show that function can be poorly represented by form and response alone. Our results suggest that the inclusion of more fundamental characteristics, in our case geology, may represent a useful path forward to capture catchment function—i.e. measures of water storage and release—in the regionalization process. Further, our findings confirm, for the first time at the landscape scale, the importance of bedrock geology in runoff generation processes in upland, headwater catchments, and suggest that this deeper dimension of catchment systems can play a significant role in controlling the fate and transport of water and solutes moving through these types of systems.

## REFERENCES


Amoozegar, A. (1989), A compact constant-head permeameter for measuring hydraulic conductivity of the vadose zone, *Soil Sci. Soc. Am. J.*, *53*, 1356-1361.

Anderson, S. P., W. E. Dietrich, D. R. Montgomery, R. Torres, M. E. Conrad, and K. Loague (1997), Subsurface flow paths in a steep, unchanneled catchment., *Water Resour. Res.*, *33*(12), 2637-2653.

Arnold, J. G., P. M. Allen, R. Muttiah, and G. Bernhardt (1995), Automated Base Flow Separation and Recession Analysis Techniques, *Ground Water*, *33*(6), 1010-1018.

Asano, Y., T. Uchida, and N. Ohte (2002), Residence times and flow paths of water in steep unchannelled catchments, Tanakami, Japan, *J. Hydrol.*, *261*(1-4), 173-192.

Baker, D. B., R. P. Richards, T. T. Loftus, and J. W. Kramer (2004), A NEW FLASHINESS INDEX: CHARACTERISTICS AND APPLICATIONS TO MIDWESTERN RIVERS AND STREAMS1, *JAWRA Journal of the American Water Resources Association*, *40*(2), 503-522.

Beven, K. J., and M. J. Kirkby (1979a), A physically-based variable contributing area model of basin hydrology, *Hydrological Sciences Bulletin*, *24*, 43-69.

Beven, K. J., and M. J. Kirkby (1979b), A physically based, variable contributing area model of basin hydrology., *Hydrological Sciences Bulletin*, *24*(1), 43-69.

Bierlmaler, F. A., and A. McKee (1989), Climatic summaries and documentation for the primary meteorological station, H.J. Andrews Experimental Forest, 1972 to 1984, *U.S. Department of Agriculture, Forest Service, Pacific Northwest Research Station*, *PNW-GTR-242*, 61.

Blöschl, G. (2001), Scaling in hydrology, *Hydrol. Processes*, *15*, 709-711.

Brutsaert, W., and J. L. Nieber (1977), REGIONALIZED DROUGHT FLOW HYDROGRAPHS FROM A MATURE GLACIATED PLATEAU, *Water Resour. Res.*, *13*(3), 637-644.

Capell, R., D. Tetzlaff, I. A. Malcolm, A. J. Hartley, and C. Soulsby (2011), Using hydrochemical tracers to conceptualise hydrological function in a larger scale catchment draining contrasting geologic provinces, *J. Hydrol.*, *408*(1-2), 164-177.

Corliss, J. F. (1973), Soil Survey, Alsea Area, Oregon, *U.S. Soil Conservation Service*, 82 pp.

DeWalle, D. R., P. J. Edwards, B. R. Swistock, R. Aravena, and R. J. Drimmie (1997), Seasonal isotope hydrology of three Appalachian forest catchments, *Hydrol. Processes*, *11*(15), 1895-1906.

Dietrich, W. E., and T. Dunne (1978), Sediment budget for a small catchment in mountainous terrain, *Z. Geomorph. Suppl.*, *29*, 191-206.

Dingman, S. L. (2002), *Physical hydrology*, 2nd ed., x, 646 p. pp., Prentice Hall, Upper Saddle River, N.J.

Dooge, J. C. I. (1986), Looking for Hydrologic Laws, *Water Resour. Res.*, *22*, 46S-58S.

Freeze, R. A., and J. A. Cherry (1979), *Groundwater*, 1st ed., 604 pp., Prentice-Hall, Englewood Cliffs, NJ.

Gleeson, T., L. Smith, N. Moosdorf, J. Hartmann, H. H. Dürr, A. H. Manning, L. P. H. van Beek, and A. M. Jellinek (2011), Mapping permeability over the surface of the Earth, *Geophys. Res. Lett.*, *38*(2), L02401.

Graham, C. B., and J. J. McDonnell (2010), Hillslope threshold response to rainfall: (2) Development and use of a macroscale model, *J. Hydrol.*, *393*(1-2), 77-93.

Hale, V. C., J. J. McDonnell, M. K. Stewart, D. K. Solomon, G. G. Ice, and R. T. Pack (this issue), Similar catchment form and rainfall-runoff responses hide radically different plumbing: (2) The role of catchment storage controls, *to be submitted to Water Resources Research*.

Haria, A. H., and P. Shand (2004), Evidence for deep subsurface flow routing in a forested upland Wales: implications for contaminant transport and stream flow generation, *Hydrol. Earth Syst. Sci.*, *8*(2), 334-344.

Harr, R. D. (1976), Hydrology of small forest stream in western Oregon, General Technical Report PNW-55*Rep.*, 15 pp, Pacific Northwest Forest and Range Experiment Station, USDA Forest Service, Portland, OR.

Harr, R. D. (1977), Water flux in soil and subsoil on a steep forested slope, *J. Hydrol.*, *33*, 37-58.

Harr, R. D. (1981), Some characteristics and consequences of snowmelt during rainfall in western, Oregon, *J. Hydrol.*, *53*, 277-304.

Harris, D. D. (1977), Hydrologic changes after logging in two small Oregon coastal watersheds, *United States Geological Survey*, *Water-Supply Paper 2037*.

Hewlett, J. D., and A. R. Hibbert (1967), Factors affecting the response of small watersheds to precipitation in humid areas, in *Forest Hydrology.*, edited by W. E. Sopper and H. W. Lull, pp. 275-291, Pergamon Press, New York.

Hjerdt, K. N., J. J. McDonnell, J. Seibert, and A. Rodhe (2004), A new topographic index to quantify downslope controls on local drainage, *Water Resour. Res.*, *40*, W05602, doi:05610.01029/02004WR003130.

Hrachowitz, M., C. Soulsby, D. Tetzlaff, J. J. C. Dawson, and I. A. Malcolm (2009), Regionalization of transit time estimates in montane catchments by integrating landscape controls, *Water Resour. Res.*, *45*.

Hrachowitz, M., C. Soulsby, D. Tetzlaff, I. A. Malcolm, and G. Schoups (2010), Gamma distribution models for transit time estimation in catchments: Physical interpretation of parameters and implications for time-variant transit time assessment, *Water Resour. Res.*, *46*.

Ice, G. G., V. C. Hale, J. Light, J. J. McDonnell, S. H. Schoenholtz, and J. D. Stednick (2007), Testing contemporary forest practices: The Alsea Watershed Study Revisited., in *Sustaining America's Forests: Proceedings of the 2007 Society of American Foresters National Convention*, edited, Society of American Foresters, Portland, OR.

Ingebritsen, S. E., D. R. Sherrod, and R. H. Mariner (1992), RATES AND PATTERNS OF GROUNDWATER-FLOW IN THE CASCADE RANGE VOLCANIC ARC, AND THE EFFECT ON SUBSURFACE TEMPERATURES, *Journal of Geophysical Research-Solid Earth*, *97*(B4), 4599-4627.

Jefferson, A., A. Nolin, S. Lewis, and C. Tague (2008), Hydrogeologic controls on streamflow sensitivity to climate variation, *Hydrol. Processes*, *22*(22), 4371-4385.

Jefferson, A., G. E. Grant, S. L. Lewis, and S. T. Lancaster (2010), Coevolution of hydrology and topography on a basalt landscape in the Oregon Cascade Range, USA, *Earth Surf. Processes Landforms*, *35*(7), 803-816.

Jenny, H. (1941), *Factors of Soil Formation*, Dover Publications, Inc., New York.

Kabeya, N., M. Katsuyama, M. Kawasaki, N. Ohte, and A. Sugimoto (2007), Estimation of mean residence times of subsurface waters using seasonal variation in deuterium excess in a small headwater catchment in Japan, *Hydrol. Processes*, *21*, 308-322.

Katsuyama, M., M. Tani, and S. Nishimoto (2010), Connection between streamwater mean residence time and bedrock groundwater recharge/discharge dynamics in weathered granite catchments, *Hydrol. Processes*, *24*(16), 2287-2299.

Kirchner, J. W. (2006), Getting the right answers for the right reasons: Linking measurements, analyses, and models to advance the science of hydrology, *Water Resour. Res.*, *42*.

Kirchner, J. W., X. Feng, and C. Neal (2000), Fractal stream chemistry and its implications for contaminant transport in catchments, *Nature*, *403*(6769), 524-527.

Kosugi, K. i., S. y. Katsura, M. Katsuyama, and M. Takahisa (2006), Water flow processes in weathered granitic bedrock and their effects on runoff generation in a small headwater catchment, *Water Resour. Res.*, *42*.

Kosugi, K. i., M. Fujimoto, S. y. Katsura, H. Kato, Y. Sando, and T. Mizuyama (2011), Localized bedrock aquifer distribution explains discharge from a headwater catchment, *Water Resour. Res.*, *47*(7), W07530.

Lohse, K. A., and W. E. Dietrich (2005), Contrasting effects of soil development on hydrological properties and flow paths, *Water Resour. Res.*, *41*(12).

Maloszewski, P., and A. Zuber (1982), Determining the turnover time of groundwater systems with the aid of environmental tracers. 1. models and their applicability, *J. Hydrol.*, *57*, 207-231.

Marcus, H., and D. E. Evenson (1961), Directional permeability in anisotropic porous media, *Water Resources Center Contrib. No. 31*, University of California, Berkeley.

Mazurkiewicz, A. B., D. G. Callery, and J. J. McDonnell (2008), Assessing the controls of the snow energy balance and water available for runoff in a rain-on-snow environment, *J. Hydrol.*, *354*(1-4), 1-14.

McDonnell, J. J., et al. (2010), How old is streamwater? Open questions in catchment transit time conceptualization, modelling and analysis, *Hydrol. Processes*, *24*(12), 1745-1754.

McGlynn, B., J. McDonnell, M. Stewart, and J. Seibert (2003), On the relationships between catchment scale and streamwater mean residence time, *Hydrol. Processes*, *17*(1), 175-181.

McGlynn, B. L., and J. J. McDonnell (2003), Quantifying the relative contributions of riparian and hillslope zones to catchment runoff, *Water Resour. Res.*, *39*(11), 1310, doi:1310.1029/2003WR002091.

McGlynn, B. L., and J. Seibert (2003), Distributed assessment of contributing area and riparian buffering along stream networks, *Water Resour. Res.*, *39*(4), 1082, doi:1010.1029/2002WR001521.

McGuire, K. J., and J. J. McDonnell (2006), A review and evaluation of catchment transit time modeling, *J. Hydrol.*, *330*(3-4), 543-563.

McGuire, K. J., and J. J. McDonnell (2010), Hydrological connectivity of hillslopes and streams: Characteristic time scales and nonlinearities, *Water Resour. Res.*, *46*.

McGuire, K. J., J. J. McDonnell, M. Weiler, C. Kendall, B. L. McGlynn, J. M. Welker, and J. Seibert (2005), The role of topography on catchment-scale water residence time, *Water Resour. Res.*, *41*(5), -.

Millares, A., M. J. Polo, and M. A. Losada (2009), The hydrological response of baseflow in fractured mountain areas, *Hydrol. Earth Syst. Sci.*, *13*(7), 1261-1271.

Montgomery, D. R., W. E. Dietrich, R. Torres, S. P. Anderson, J. T. Heffner, and K. Loague (1997), Hydrologic response of a steep, unchanneled valley to natural and applied rainfall, *Water Resour. Res.*, *33*(1), 91-109.

Olden, J. D., and N. L. Poff (2003), Redundancy and the choice of hydrologic indices for characterizing streamflow regimes, *River Res. Appl.*, *19*(2), 101-121.

Onda, Y., M. Tsujimura, J.-i. Fujihara, and J. Ito (2006), Runoff generation mechanisms in high-relief mountainous watersheds with different underlying geology, *J. Hydrol.*, *331*(3-4), 659-673.

Oudin, L., A. Kay, V. Andréassian, and C. Perrin (2010), Are seemingly physically similar catchments truly hydrologically similar?, *Water Resour. Res.*, *46*(11), W11558.

Peck, D. V., A. B. Griggs, H. G. Schlicker, F. G. Wells, and H. M. Dole (1964), Geology of the central and nothern parts of the western Cascade Range in Oregon, *U.S. Geological Survey Professional Paper 449*, 56.

Ranken, D. W. (1974), Hydrologic properties of soil and subsoil on a steep, forested slope, M.S. thesis, 114 pp, Oregon State University, Corvallis.

Rodgers, P., C. Soulsby, and D. Tetzlaff (2005), Using stable isotope tracers to assess hydrological flow paths, residence times and landscape influences in a nested mesoscale catchment, *Hydrol. Earth Syst. Sci.*, *9*, 139-155.

Roering, J. J., J. W. Kirchner, and W. E. Dietrich (2005), Characterizing structural and lithologic controls on deep-seated landsliding: Implications for topographic relief and landscape evolution in the Oregon Coast Range, USA, *Geological Society of America Bulletin*, *117*(5-6), 654-668.

Rupp, D. E., and R. A. Woods (2008), Increased flexibility in base flow modelling using a power law transmissivity profile, *Hydrol. Processes*, *22*(14), 2667-2671.

Saar, M. O., and M. Manga (2004), Depth dependence of permeability in the Oregon Cascades inferred from hydrogeologic, thermal, seismic, and magmatic modeling constraints, *J. Geophys. Res.*, *109*(B4), B04204.

Sawicz, K., T. Wagener, M. Sivapalan, P. A. Troch, and G. Carrillo (2011), Catchment classification: empirical analysis of hydrologic similarity based on catchment function in the eastern USA, *Hydrol. Earth Syst. Sci.*, *15*(9), 2895-2911.

Schoups, G., and J. A. Vrugt (2010), A formal likelihood function for parameter and predictive inference of hydrologic models with correlated, heteroscedastic, and non-Gaussian errors, *Water Resour. Res.*, *46*.

Sivapalan, M. (2003), Process complexity at hillslope scale, process simplicity at the watershed scale: is there a connection?, *Hydrol. Processes*, *17*(5), 1037-1041.

Sivapalan, M. (2005), Pattern, process, and function: Elements of a unified theory of hydrology at the catchment scale, in *Encyclopedia of Hydrological Sciences*, edited by M. G. Anderson, pp. 193-219, John Wiley & Sons, Ltd.

Snavely, P. D., H. C. Wagner, and N. S. MacLeod (1964), Rythmic-bedded eugeosynclinal deposits of the Tyee Formation, Oregon Coast Range, in *Symposium on cyclic sedimentation: Kansas Geological Survey, Bulletin 169, pp. 461-480*, edited by D. F. Merriam.

Soulsby, C., D. Tetzlaff, and M. Hrachowitz (2009), Tracers and transit times: windows for viewing catchment scale storage?, *Hydrol. Processes*, *23*(24), 3503-3507.

Soulsby, C., D. Tetzlaff, and M. Hrachowitz (2010), Are transit times useful process-based tools for flow prediction and classification in ungauged basins in montane regions?, *Hydrol. Processes*, *24*(12), 1685-1696.

Soulsby, C., R. Malcolm, R. C. Ferrier, R. C. Helliwell, and A. Jenkins (2000), Isotope hydrology of the Allt a'Mharcaidh catchment, Cairngorms, Scotland: implications for hydrological pathways and residence times, *Hydrol. Processes*, *14*, 747-762.

Soulsby, C., D. Tetzlaff, P. Rodgers, S. Dunn, and S. Waldron (2006), Runoff processes, stream water residence times and controlling landscape characteristics in a mesoscale catchment: An initial evaluation, *J. Hydrol.*, *325*(1-4), 197-221.

Soulsby, C., D. Tetzlaff, N. van den Bedem, I. A. Malcolm, P. J. Bacon, and A. F. Youngson (2007), Inferring groundwater influences on surface water in montane catchments from hydrochemical surveys of springs and streamwaters, *J. Hydrol.*, *333*(2-4), 199-213.

Stanfield, B. J., J. C. Bliss, and T. A. Spies (2002), Land ownership and landscape structure: a spatial analysis of sixty-six Oregon (USA) Coast Range watersheds, *Landscape Ecology*, *17*(8), 685-697.

Stewart, M. K., and J. J. McDonnell (1991), Modeling base flow soil water residence times from deuterium concentrations, *Water Resour. Res.*, *27*(10), 2681-2693.

Stewart, M. K., U. Morgenstern, and J. J. McDonnell (2010), Truncation of stream residence time: how the use of stable isotopes has skewed our concept of streamwater age and origin, *Hydrol. Processes*, *24*(12), 1646-1659.

Swanson, F. J., and M. E. James (1975), Geology and geomorphology of the H.J. Andrews Experimental Forest, western Cascades, Oregon.*Rep. Res. Pap. PNW-188*, 14 pp, U.S. Department of Agriculture, Forest Service, Pacific Northwest Forest and Range Experiment Station, Portland, OR.

Tague, C., and G. E. Grant (2004), A geological framework for interpreting the low-flow regimes of Cascade streams, Willamette River Basin, Oregon, *Water Resour. Res.*, *40*(W04303), doi:10.1029/2003WR002629.

Tarboton, D. G. (1997), A new method for the determination of flow directions and upslope areas in grid digital elevation models, *Water Resour. Res.*, *33*, 309-319.

Tetzlaff, D., J. Seibert, and C. Soulsby (2009a), Inter-catchment comparison to assess the influence of topography and soils on catchment transit times in a geomorphic province; the Cairngorm mountains, Scotland, *Hydrol. Processes*, *23*(13), 1874-1886.

Tetzlaff, D., S. K. Carey, H. Laudon, and K. McGuire (2010), Catchment processes and heterogeneity at multiple scales—benchmarking observations, conceptualization and prediction, *Hydrol. Processes*, *24*(16), 2203-2208.

Tetzlaff, D., J. Seibert, K. J. McGuire, H. Laudon, D. A. Burns, S. M. Dunn, and C. Soulsby (2009b), How does landscape structure influence catchment transit time across different geomorphic provinces?, *Hydrol. Processes*, *23*(6), 945-953.

Tetzlaff, D., J. J. McDonnell, S. Uhlenbrook, K. J. McGuire, P. W. Bogaart, F. Naef, A. J. Baird, S. M. Dunn, and C. Soulsby (2008), Conceptualizing catchment processes: simply too complex?, *Hydrol. Processes*, *22*(11), 1727-1730.

Thorson, T. D., S.A. Bryce, D.A. Lammers, A.J. Woods, J.M. Omernik, J. Kagan, D.E. Pater, and J.A. Comstock (2003), Ecoregions of Oregon, *U.S. Geological Survey, Reston, VA.*

Torres, R., W. E. Dietrich, D. R. Montgomery, S. P. Anderson, and K. Loague (1998), Unsaturated zone processes and the hydrologic response of a steep, unchanneled catchment, *Water Resour. Res.*, *34*(8), 1865-1879.

Uchida, T., J. J. McDonnell, and Y. Asano (2006), Functional intercomparison of hillslopes and small catchments by examining water source, flowpath and mean residence time, *J. Hydrol.*, *327*, 627-642.

Wagener, T., M. Sivapalan, P. Troch, and R. Woods (2007), Catchment classification and hydrologic similarity, *Geography Compass*, *1*.

Wilson, C. J., and W. E. Dietrich (1987), The contribution of bedrock groundwater flow to storm runoff and high pore pressure development in hollows, in *Erosion and Sedimentation in the Pacific Rim*, edited, IAHS Publication No. 165, Corvallis, OR.

Yadav, M., T. Wagener, and H. Gupta (2007), Regionalization of constraints on expected watershed response behavior for improved predictions in ungauged basins, *Adv. Water Resour.*, *30*(8), 1756-1774.

Table 2.1. Terrain and hydrologic metrics calculated for the sedimentary Oregon Coast Range (NB-12 and NB-86) and volcanic Western Cascades research catchments (WS10 and WS01).

| | *NB-12* | *NB-86* | *WS10* | *WS01* |
|---|---|---|---|---|
| ***Terrain Metrics*** | | | | |
| $A_c$ [ha] | 11.9 | 85.7 | 10.2 | 95.9 |
| $E_{min}$ [m] | 207 | 132 | 462 | 440 |
| $E_{rng}$ [m] | 162 | 237 | 226 | 570 |
| $DD$ [km km$^{-2}$] | 2.53 | 3.70 | 3.01 | 3.72 |
| $A$-$P$ [km$^2$ km$^{-1}$] | 0.07 | 0.15 | 0.06 | 0.175 |
| Median $S$ [%] | 51 | 34 | 66 | 68 |
| Median $L$ [m] | 134 | 98 | 137 | 139 |
| Median $G$ [m m$^{-1}$] | 0.47 | 0.34 | 0.63 | 0.61 |
| Median $SCA$ [ha] | 4.1 | 12.3 | 7.2 | 10.3 |
| Median $TWI$ | 6.30 | 6.58 | 4.27 | 4.41 |
| Median $DSI_5$ | 0.45 | 0.32 | 0.52 | 0.57 |
| | | | | |
| ***Hydrologic Metrics*** | | | | |
| $MAF$ [mm d$^{-1}$] | 4.46 | 4.24 | 4.04 | 3.71 |
| $MAPF$ [mm d$^{-1}$] | 65.38 | 62.79 | 62.39 | 60.52 |
| $MALF$ [mm d$^{-1}$] | 0.32 | 0.04 | 0.02 | 0.03 |
| $CV_Q$ | 1.85 | 1.76 | 1.99 | 1.91 |
| $R_{QP}$ | 0.95 | 0.78 | 0.65 | 0.59 |
| $BFI$ | 0.64 | 0.72 | 0.62 | 0.68 |
| $FDC_{33-66}$ | 3.54 | 3.41 | 4.21 | 3.61 |
| $FI_{RB}$ | 0.42 | 0.33 | 0.45 | 0.40 |
| $B$ | 1.64 | 1.10 | 1.23 | 1.20 |
| | | | | |
| ***Mean Transit Time* [y]** | 5.0 | 4.0 | 1.2 | 1.3 |

Table 2.2. Maximum likelihood estimates (mle) for the alpha ($\alpha$) and beta ($\beta$) parameters of the gamma transit time distribution model, mean transit times (MTT), uncertainties, and Nash-Sutcliffe efficiencies (NSE) for catchments in the Drift Creek basin in the Oregon Coast Range.

| Location | $\alpha_{mle}$ | $\alpha_{10/90\%}$ | $\beta_{mle}$ [y] | $\beta_{10/90\%}$ [y] | $MTT_{mle}$ [y] | $MTT_{10/90\%}$ [y] | $NSE_{mle}$ |
|---|---|---|---|---|---|---|---|
| NB-12 | 1.44 | (1.01/1.46) | 3.5 | (2.9/8.2) | 5.0 | (4.0/8.7) | 0.34 |
| NB-35 | 1.48 | (1.30/1.49) | 2.5 | (2.2/3.3) | 3.7 | (3.2/4.5) | 0.38 |
| NB-86 | 1.44 | (1.28/1.49) | 2.8 | (2.4/3.7) | 4.0 | (3.5/4.9) | 0.47 |
| FC-210 | 1.33 | (1.30/1.49) | 4.7 | (3.4/7.5) | 6.3 | (5.0/10.1) | 0.30 |
| DC-315 | 1.32 | (0.98/1.46) | 3.5 | (2.6/12.5) | 4.7 | (3.7/11.6) | 0.23 |
| MC-1881 | 1.37 | (1.04/1.47) | 4.0 | (3.0/13.7) | 5.5 | (4.3/14.0) | 0.33 |
| DR-5373 | 1.37 | (1.22/1.50) | 7.6 | (5.6/10.8) | 10.4 | (8.3/15.7) | 0.30 |
| DR-8643 | 1.49 | (1.42/1.51) | 6.8 | (5.3/12.8) | 10.2 | (7.8/18.3) | 0.46 |

Table 2.3. Catchment area ($A_c$), mean transit times (MTT), uncertainties, and Nash-Sutcliffe efficiencies (NSE) estimated using the exponential transit time distribution for catchments at the HJ Andrews Experimental Forest in the Western Cascades range of Oregon (data sourced from *McGuire et al.* [2005]).

| Location | $A_c$ [ha] | MTT [y] | MTT +/-2$\sigma_p$ [y] | NSE |
|---|---|---|---|---|
| WS02 | 60.1 | 2.2 | (1.6/2.8) | 0.45 |
| WS03 | 101.1 | 1.3 | (1.0/1.6) | 0.48 |
| WS08 | 21.4 | 3.3 | (2.0/4.6) | 0.40 |
| WS09 | 8.5 | 0.8 | (0.6/1.0) | 0.46 |
| WS10 | 10.2 | 1.2 | (0.9/1.5) | 0.49 |
| MACK | 581 | 2.0 | (1.5/2.5) | 0.54 |
| LOOK | 6242 | 2.0 | (1.0/3.0) | 0.32 |

Table 2.4. Pearson's correlation coefficients and associated p-values between mean transit time (MTT) and catchment attributes for catchments within the Drift Creek basin in the Oregon Coast Range and the HJ Andrews Experimental Forest (HJA) in the Western Cascades range of Oregon.

| Catchment Attribute | Upper Drift Creek Basin | | HJA Experimental Forest | |
|---|---|---|---|---|
| | Correlation Coefficient | p-value | Correlation Coefficient | p-value |
| $A_c$ | 0.91 | <0.01 | 0.00 | 0.99 |
| log($A_c$) | 0.82 | 0.01 | -0.03 | 0.95 |
| *DD* | 0.17 | 0.70 | 0.69 | 0.13 |
| *A-P* | 0.91 | <0.01 | -0.01 | 0.99 |
| *S* | -0.15 | 0.73 | -0.86 | 0.03 |
| *L* | 0.58 | 0.13 | 0.87 | 0.03 |
| *G* | 0.31 | 0.45 | -0.80 | 0.05 |
| *L/G* | 0.13 | 0.75 | 0.95 | <0.01 |
| *SCA* | -0.15 | 0.73 | -0.54 | 0.27 |
| *TWI* | -0.12 | 0.79 | 0.65 | 0.16 |
| $DSI_5$ | -0.12 | 0.79 | 0.65 | 0.16 |

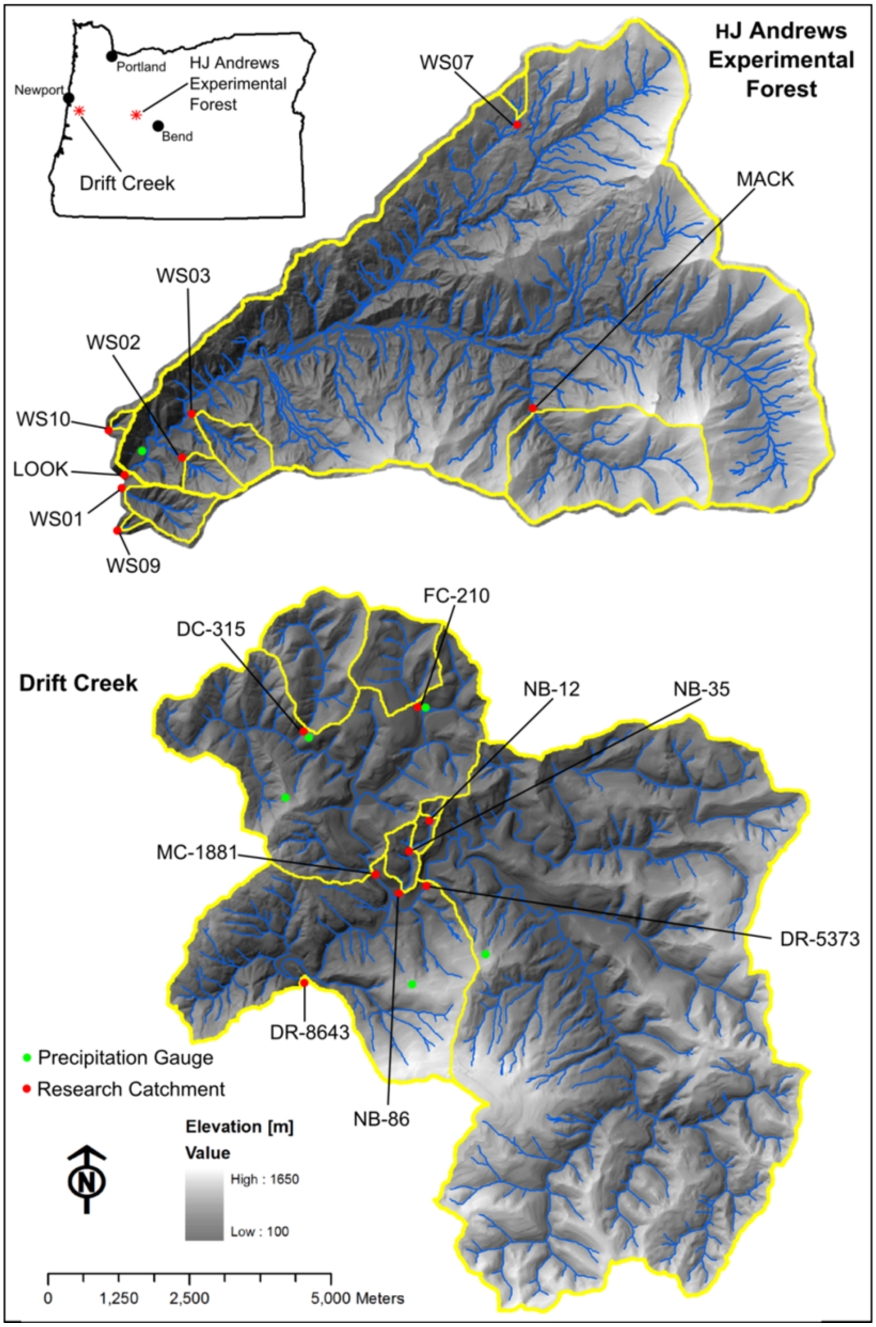


Figure 2.1. Map of the Drift Creek and HJ Andrews Experimental Forest with inlay showing their general locations within the state of Oregon. The research catchments and precipitation gauges used in this study are labeled.

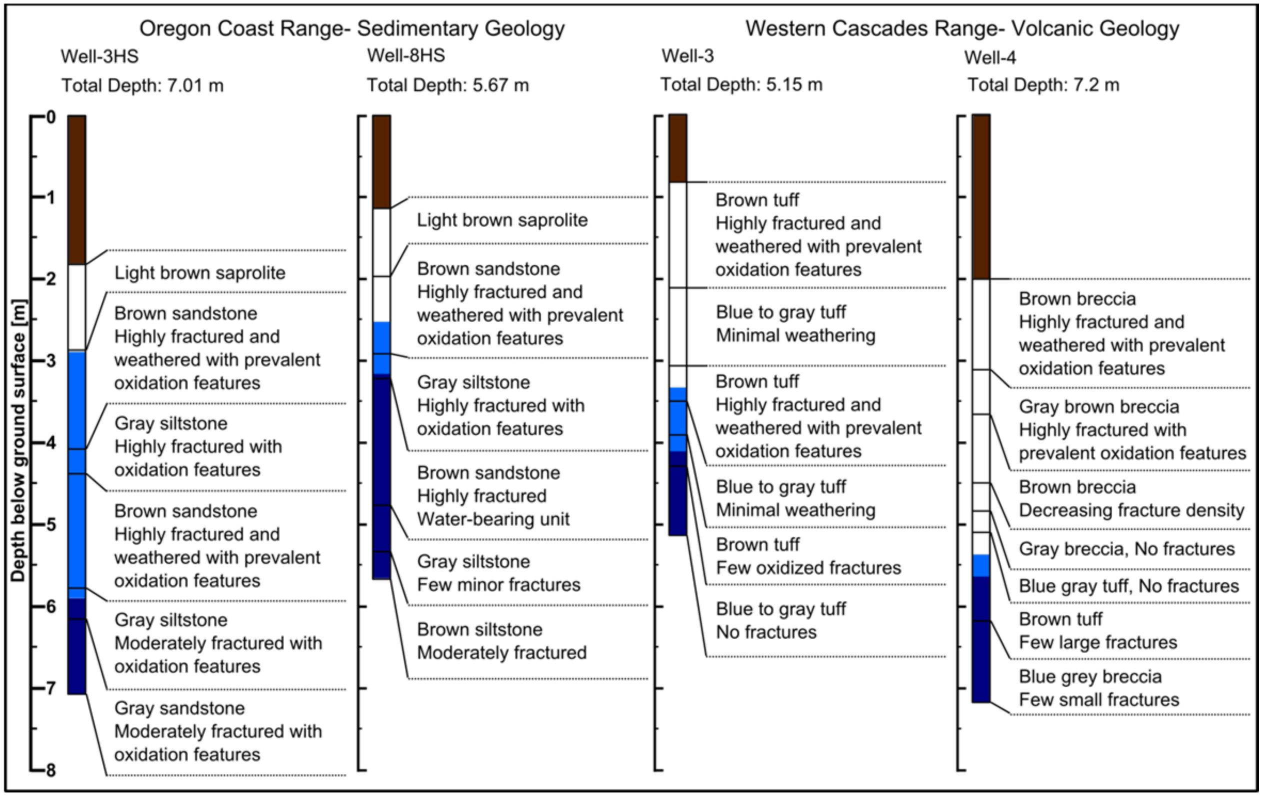


Figure 2.2. Boring logs from shallow bedrock wells installed within NB-12 in the Drift Creek basin of the Oregon Coast Range (Well-3HS and Well-8HS) and WS10 in the HJ Andrews Experimental Forest in the Western Cascades range (Well-3 and Well-4). Brown denotes soil, dark blue shows the minimum water level, and light blue shows the range of measured water level fluctuations.

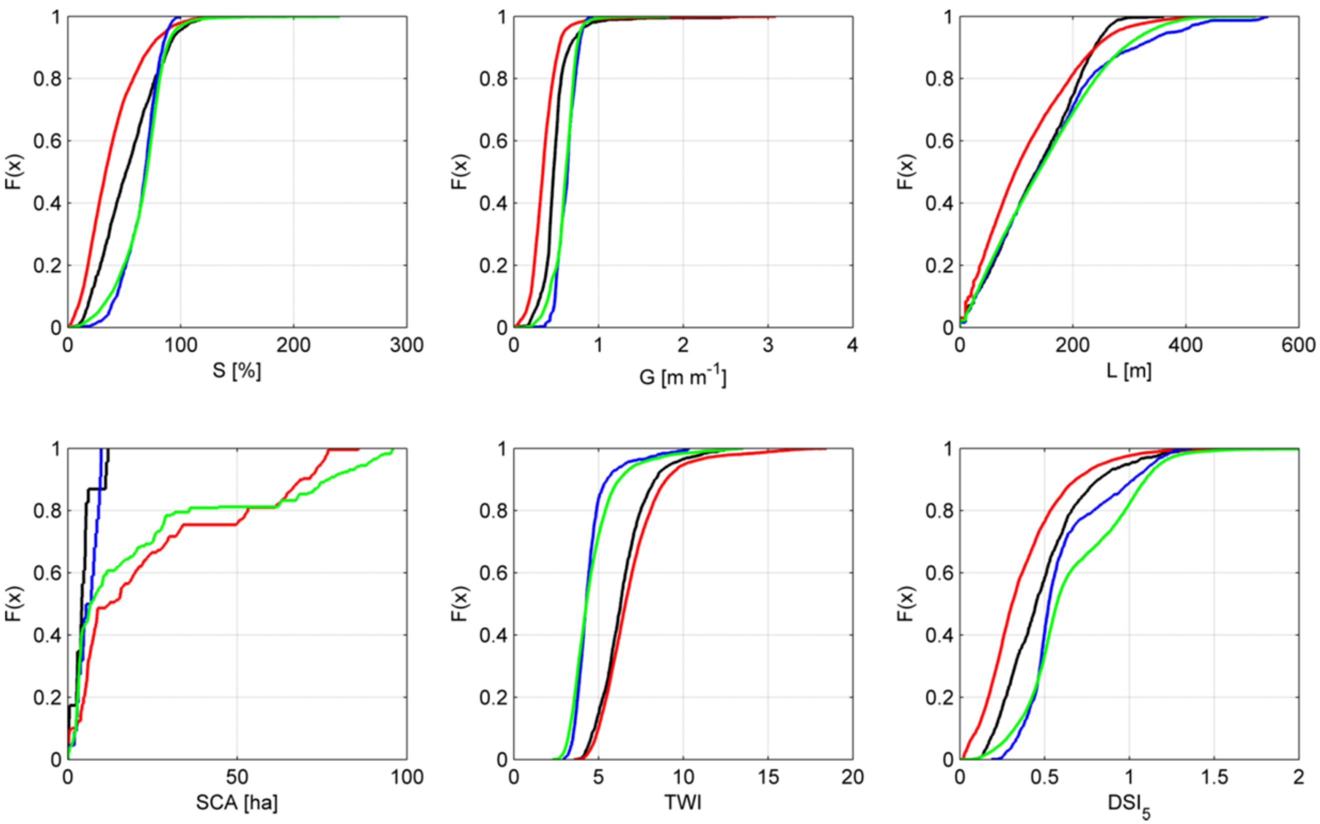


Figure 2.3. Empirical cumulative distribution plots of slope (a), flowpath gradient (b), flowpath length (c), subcatchment area (d), topographic wetness index (e), and downslope index (f) for NB-12 (black), NB-86 (red), WS10 (blue), and WS01 (green).

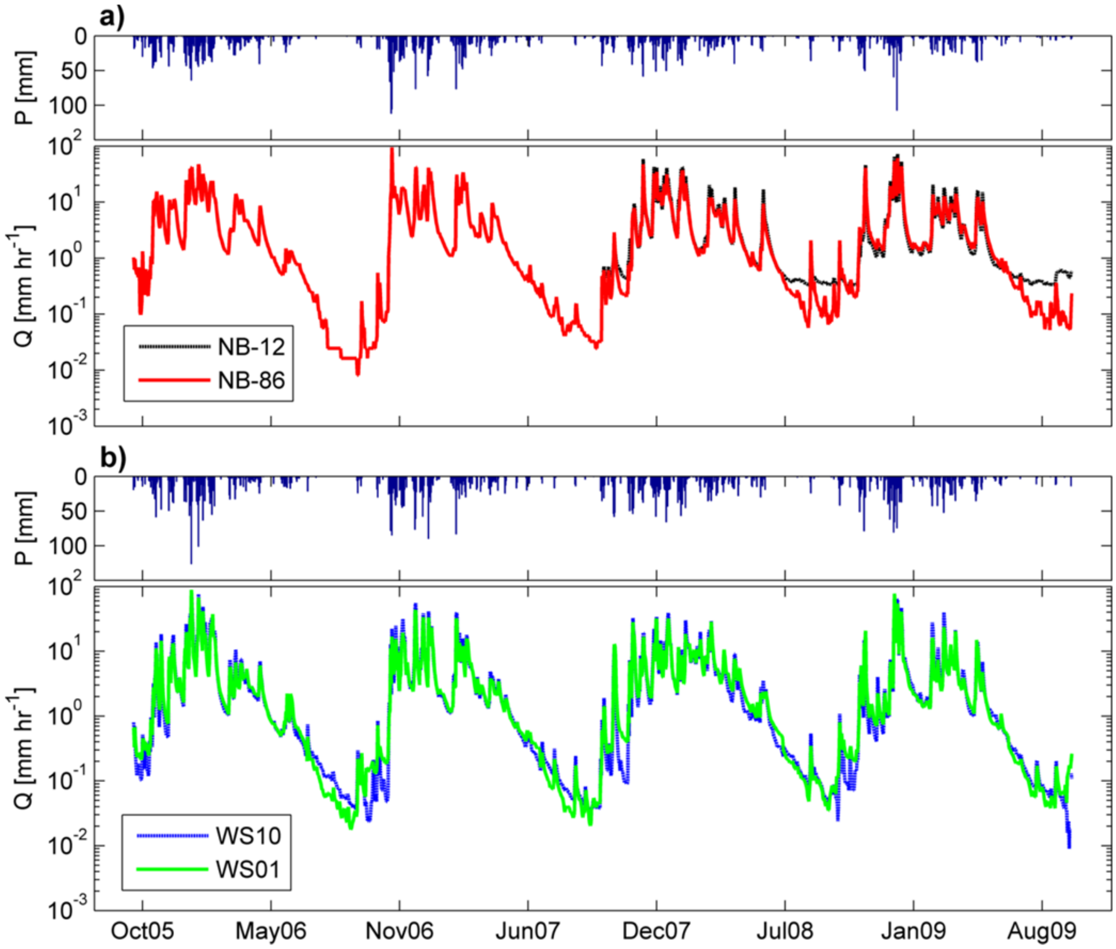


Figure 2.4. Plot of mean daily discharge and daily precipitation measured at NB-12 (black), NB-86 (red), WS10 (blue), and WS01 (green) during the intercomparison period.

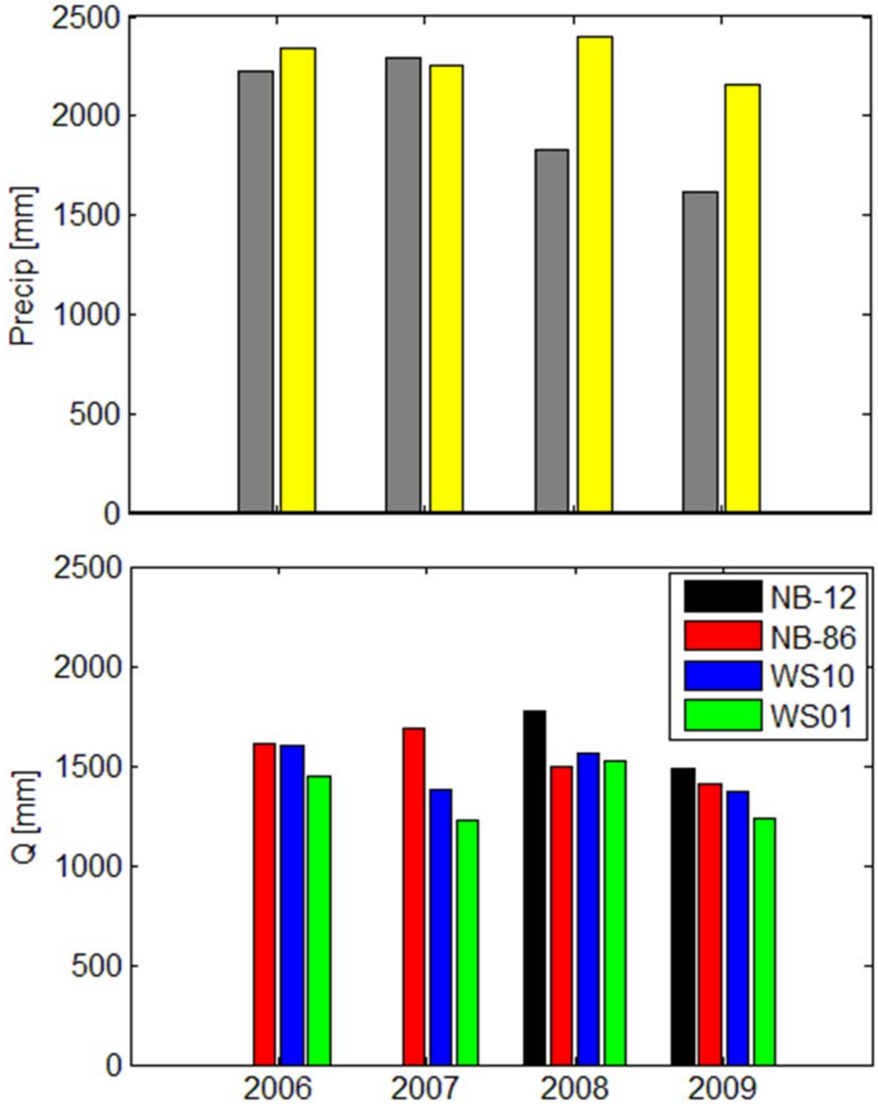


Figure 2.5. Annual precipitation (a; gray is NB-12 and NB-86, yellow is WS10 and WS01) and discharge (b) during intercomparison period. Discharge was not measured at NB-12 during 2006 and 2007.

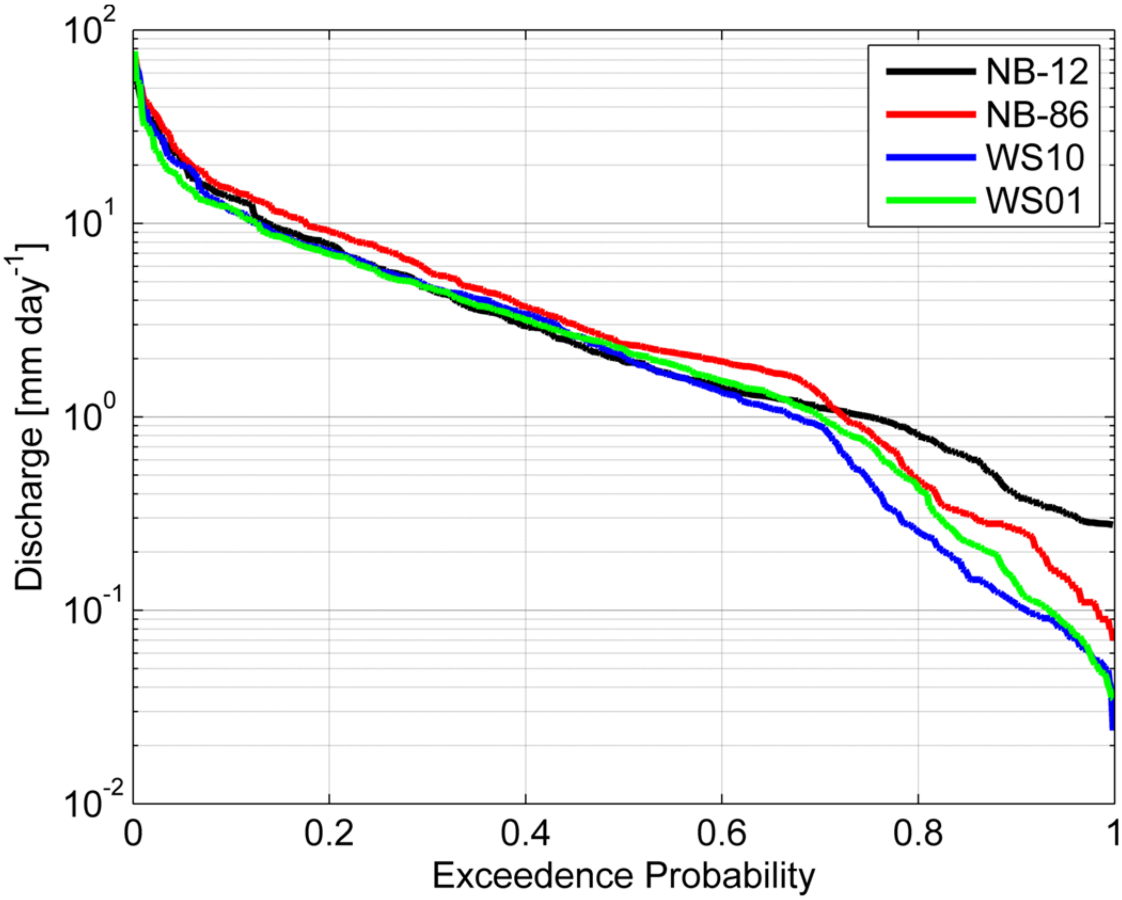


Figure 2.6. Flow duration curves for NB-12, NB-86, WS10, and WS01.

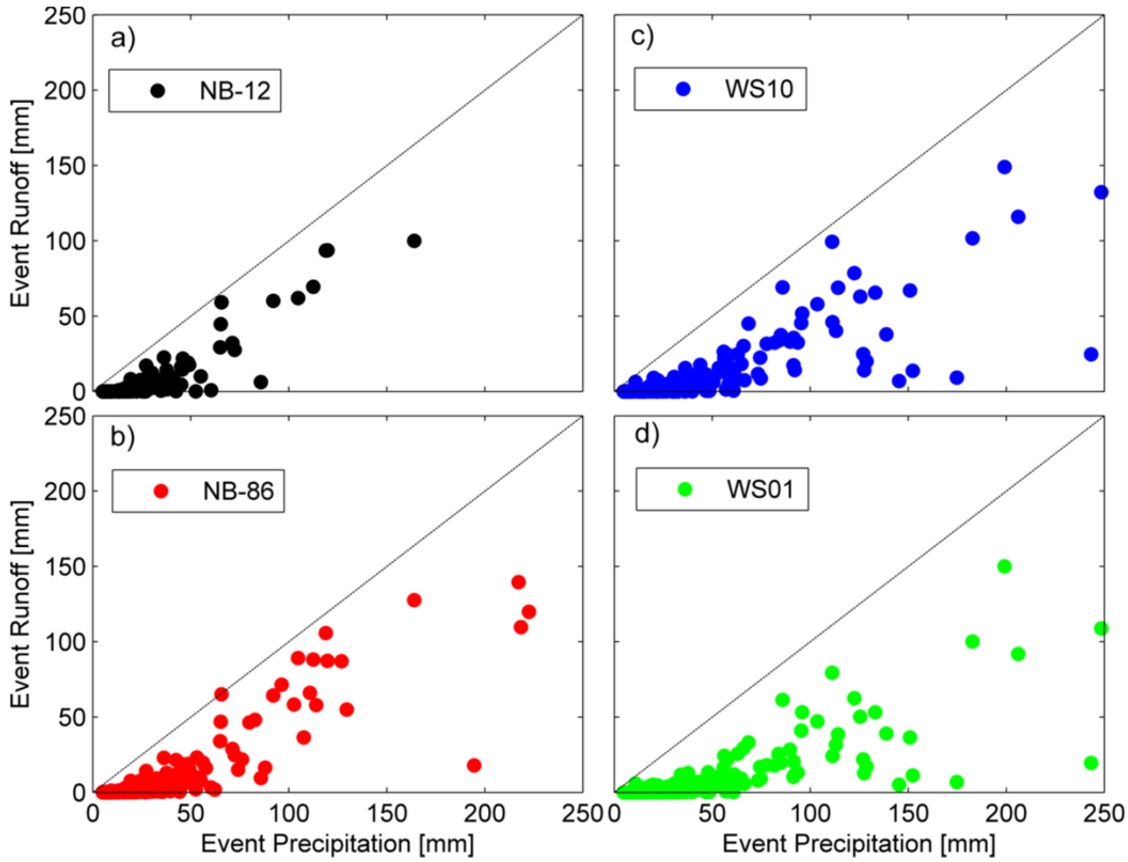


Figure 2.7. Event runoff ($Q_{evt}$) versus event precipitation ($P_{evt}$) for the Drift Creek (a=NB-12; b=NB-86) and HJA ( c=WS10; d=WS01).

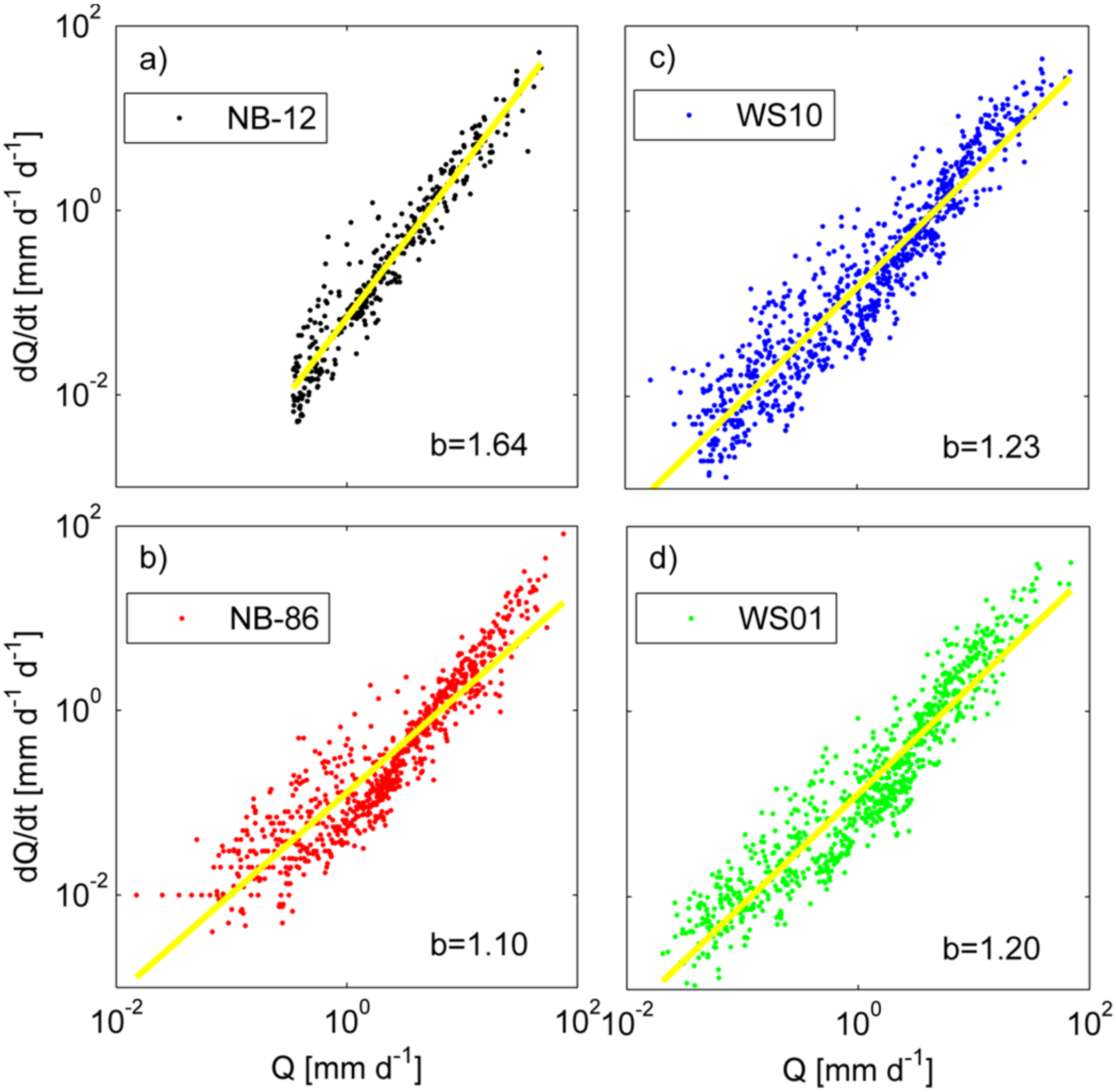


Figure 2.8. Plot of dQ/dt-Q relationships for NB-12 (a), NB-86 (b), WS10 (c), and WS01 (d).

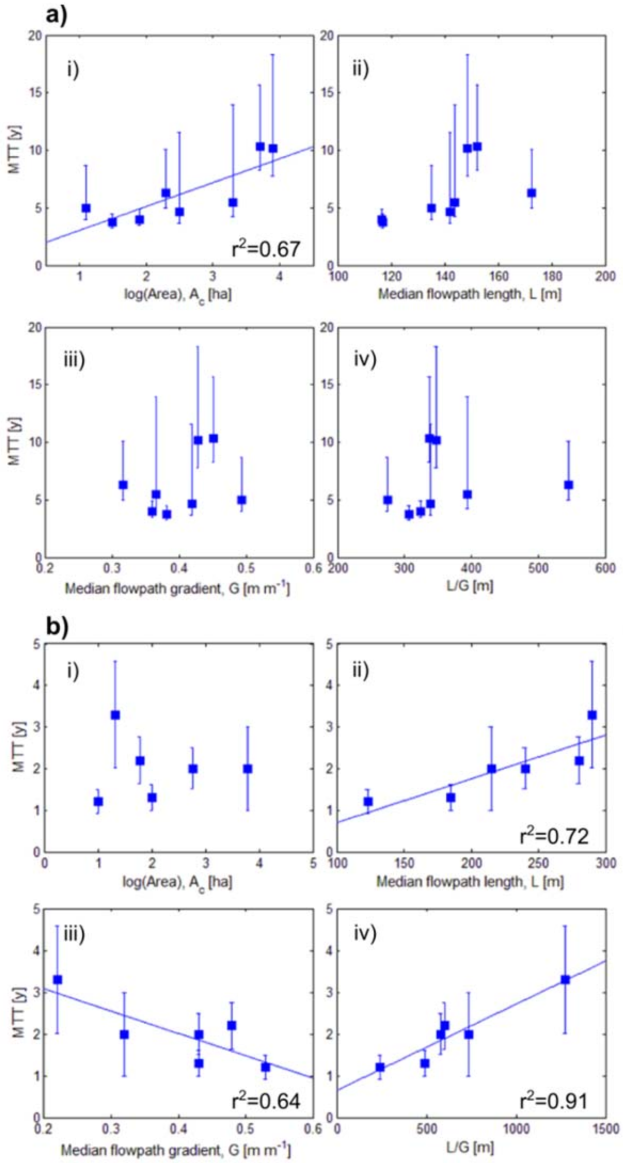


Figure 2.9. Mean transit time as a function of the logarithm of catchment area (i), median flowpath median flowpath length (ii), median flowpath gradient (iii), and the ratio of median flowpath length and flowpath gradient (iv) for the sedimentary Drift Creek catchments (a) and the volcanic HJ Andrews catchments (b).

**Similar catchment forms and rainfall-runoff responses hide radically different plumbing: (2) A bottom-up catchment storage investigation**

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## INTRODUCTION

Despite well-established theory relating the amount of water stored in a catchment to stream discharge [*Botter et al.*, 2009; *Brutsaert and Nieber*, 1977; *Kirchner*, 2009], process understanding of subsurface water storage remains a poorly understood component of catchment hydrology [*McNamara et al.*, 2011]. In early conceptualizations of subsurface storm runoff generation processes [*Hewlett and Hibbert*, 1967; *Weyman*, 1973; *Whipkey*, 1965], storage was generally considered to be limited to the soil pore space as the underlying bedrock was assumed impermeable (see *Bonell* [1993] for review). However, over the past two decades, researchers have provided unequivocal evidence linking deeper storage components to runoff generation in steep mountainous catchments [including *Anderson et al.*, 1997b; *Asano et al.*, 2002; *Haria and Shand*, 2004; *Kosugi et al.*, 2006; *Kosugi et al.*, 2011; *Millares et al.*, 2009; *Soulsby et al.*, 2007; *Wilson and Dietrich*, 1987]. These storage components can lead to very long residence times of water in the subsurface both during and between events and can skew our understanding of streamflow sources, flowpaths, and transit times [*Stewart et al.*, 2010].

Differences in subsurface storage properties have been observed and related to runoff dynamics [*Ali et al.*, 2011; *Kosugi et al.*, 2006; *Oswald et al.*, 2011; *Spence*, 2010] and solute concentrations [*Anderson et al.*, 1997a; *Birkel et al.*, 2011b; *Haria and Shand*, 2004; *Soulsby et al.*, 2007]. However, storage is an inherently difficult characteristic to study because it is largely unobservable with current technology [*Soulsby et al.*, 2009]. Despite recent advancements in ground-based geophysics [*Robinson et al.*, 2008] and gravity remote sensing [*Reager and Famiglietti*, 2009], the total storage volume of a catchment is virtually unknowable because the overall control volume is ill-defined. Even when the control volume is assumed, the internal pore volume and pore connectivity are still poorly characterized by current measurement technologies [*Rinaldo et al.*, 2011]. Although recent estimates of storage have been attempted by extrapolating point-based measurements [*McNamara et al.*, 2011] and via catchment-based mass balance [*Sayama et al.*, 2011], one of the best proxies for characterizing catchment storage is streamwater mean transit time [*Birkel et al.*, 2011a; *Fenicia et al.*, 2010; *Katsuyama et al.*, 2010; *Uchida et al.*, 2006]. Water transit time is defined as the time elapsed between the point of entry and point of exit for any given parcel of water flowing through the catchment system [*McDonnell et al.*, 2010]. Conservative, naturally occurring stable isotopes of water are often used to define such water entry and exit. Streamwater mean transit time (MTT) is the average of the individual travel times associated with the various water flowpaths within the catchment system that contribute to streamflow at a given instant [*McGuire and McDonnell*, 2006]. In its simplest form, MTT is catchment storage divided by discharge. This important link allows storage to be estimated when stream discharge and MTT are known [*Soulsby et al.*, 2009]. Nevertheless, very few studies have used MTT to estimate catchment water storage.

*Hale and McDonnell* [this issue] used an MTT intercomparison analysis to show that catchment storage differences may be masked by similarities in climate, vegetation, morphology, and hydrologic flow regimes. Their findings revealed that despite striking similarities in hydrometric and hydromorphic properties, catchments in the Oregon Coast Range with permeable sedimentary geology (Drift Creek) had distinctly longer MTTs (mean MTT was 6.2 yrs, n=8) than catchments in the nearby (<140 km) Western Cascades Range with tight volcanic geology (HJA; mean MTT was 1.8 yrs, n=7). More importantly, the scaling relationships between MTT and the overlying landscape were opposite; MTTs at the permeable sedimentary catchments scaled best with catchment area while MTTs at the volcanic catchments were highly correlated to indices of catchment topography (flowpath gradient and length) and showed no correlation with catchment area.

So how might catchment storage control MTT ranges and scaling? *Sayama and McDonnell* [2009] showed the effect of catchment storage volume on MTT ranges using a modeling experiment where catchment topography, soils, and climate of two well-studied catchments (HJ Andrews WS10 (Oregon, USA) and Maimai M8 (New Zealand)) were interchanged to investigate their influence on streamwater MTT. They found that greater soil depth, and hence greater storage, led to longer transit times regardless of which topographic configuration or climate type was used. However, in catchments with complex subsurface features such as fractured bedrock, the relationship between storage and MTT is not as straight-forward [*Rinaldo et al.*, 2011]. In situations where multiple compartmentalized bedrock aquifers actively and independently contribute to streamflow—as has been observed in Japan [*Kosugi et al.*, 2011] and Wales [*Haria and Shand*, 2004]—stream water MTT will be the integration of the unique transit time distribution of each storage component. Therefore, the age and relative contribution of each storage component contributing to streamflow sets the streamwater MTT [*Botter et al.*, 2010]. Further, if the relative contribution of the individual storage components to streamflow changes with storage volume, then streamwater MTT will vary as catchment storage varies, as shown by *Niemi* [1978]. To date, no studies that we are aware of have assessed the role of catchment storage on the scaling relationships of MTT. However, *Hale and McDonnell* [this issue] hypothesized that the primary control on the MTT ranges and scaling relations they observed was the difference between the subsurface permeability contrasts resulting from the geology type of their research catchments.

Water storage and flow in permeable bedrock can lead to long streamwater MTTs. Mean transit times on the order of 30 to 145 yrs have been estimated at spring-fed streams or streams receiving flow contributions from multiple, permeable bedrock aquifers in New Zealand catchments [*Morgenstern*, 2007; *Morgenstern et al.*, 2005]. However, when one or more of the water stores contributing to streamflow contains relatively old water, a methodological problem in MTT analysis arises, whereby the calculated MTT becomes dependent on the type of tracer used in the convolution model [Stewart et al., 2010]. The response function used in the lumped-parameter convolution approach causes the modeled output of a conservative tracer (i.e. commonly used stable isotopes, oxygen-18 and deuterium, or chloride) to become flat at MTTs greater than 4 or 5 years, depending on the function used to describe the transit time distribution [*McGuire and McDonnell*, 2006]. Thus, the transit time distributions for catchments having more tracer mass within their tails (i.e. old water components) are mischaracterized when conservative tracers are used. Tritium ($^3H$) overcomes this limitation when used as a MTT tracer because the decay function that accounts for the half-life of this radioisotope provides additional constraint to the model so that the problem of a flattened output signal is avoided [*Maloszewski and Zuber*, 1982]. Indeed, *Stewart et al.* [2010] recently showed that stable isotope-based MTT estimates can be truncated relative to $^3H$-based estimates when streamwater includes old water components. This leads to an underestimation of both the importance and the size of any older storage components when using MTTs calculated with conservative tracers.

Here we provide a mechanistic assessment of catchment storage controls on streamwater MTTs that addresses the grand challenge of conceptualizing hydrologic response and transport in catchments with complex subsurface storage and old water contributions. We address fundamental questions linking bedrock permeability to catchment storage and MTTs and attempt to answer why the MTT ranges and scaling relations are so different between the tight volcanic bedrock catchments of the Western Cascades and the more-permeable sedimentary catchments of the Oregon Coast Range [following *Hale and McDonnell*, this issue]. We focus this bottom-up analysis on the permeable Coast Range site as the process reasons for the scaling relations at the HJA catchments are already well established [*McGuire and*

*McDonnell*, 2010; *McGuire et al.*, 2005] and consistent with the many MTT scaling studies that have shown catchment topography to be a primary control on MTT in steep terrain [*Broxton et al.*, 2009; *Hrachowitz et al.*, 2009; *Rodgers et al.*, 2005; *Tetzlaff et al.*, 2009] and because little is known about the storage controls on streamwater MTT in catchments with permeable geology.

We investigate catchment storage at the permeable geology catchments of *Hale and McDonnell* [this issue]using both water balance methods and MTT-based estimates of storage volumes. Water balance storage estimations, referred to as "dynamic storage", are determined by the fluxes of water into and out of the catchment over a given period of time [*Sayama et al.*, 2011]. Since MTT-based estimates of storage reference the subsurface volume that the tracer mixes with, which is likely a different quantity that the total volume of stored water, it is referred to as "passive storage" [*Birkel et al.*, 2011a]. To ensure that our MTT-based storage estimates are not biased by the MTT truncation reported *by Stewart et al.* [2010], we use $^{3}H$-based MTT estimates together with our deuterium-based MTT estimates from *Hale and McDonnell* [this issue] to estimate passive storage. As a result of new, ultra-low level tritium ($^{3}H$) measurement capabilities [*Morgenstern and Taylor*, 2009] and a well-characterized $^{3}H$ input function for our study region [*Bob Michel*, pers. comm., 2011], we are able to report, for the first time in the Northern Hemisphere, $^{3}H$-based MTT estimates from single stream water samples. In addition, we use a network of shallow and deep groundwater wells to interrogate and attempt to delineate different subsurface storage zones with the catchment. We use $^{3}H/^{3}He$ groundwater dating to help develop our conceptual model of how subsurface storage controls streamwater MTTs. Although groundwater dating has been proven to be an invaluable tool for understanding groundwater systems, it has rarely been applied to conceptualize bedrock groundwater in mountainous catchments[*Manning and Caine*, 2007].

Overall, we combine our MTT-based storage analyses and $^{3}H/^{3}He$ groundwater dating with water balance-based storage estimates, borehole characterizations, and groundwater dynamics to answer the following specific questions:

1) What are the different subsurface storage components in the sedimentary Oregon Coast Range catchments?

2) How old is the water in the delineated subsurface storage zones?

3) How do the different storage components interact and combine to set the streamwater MTTs and control the MTT-catchment area relationship?

4) Do dynamic storage and passive storage estimates provide the same information in a permeable bedrock system?

Lastly, we relate these findings to the intercomparison with the HJ Andrews catchment outlined in *Hale and McDonnell* [this issue] to form an overall conceptualization of how bedrock permeability controls subsurface storage, MTTs, and MTT scaling relationships.

**STUDY SITE**

We use the Needle Branch catchment located in the Drift Creek drainage of the Alsea River in the central Coast Range of Oregon (Figure 3.1; 44.5°N 123.9°W) to investigate the role of catchment storage in controlling MTTs. Needle Branch lies within a highly-dissected mountainous area, characterized by short, steep slopes which give rise to medium- to high-gradient streams [*Thorson*, 2003]. The catchment is forested with an even-age stand (approximately 40-yr-old) dominated by Douglas-fir (*Pseudotsuga menziesii*) with red alder (*Alnus rubra*) occurring predominantly along the riparian corridor and occasionally on the hillslopes. Mean annual precipitation for

Needle Branch is 2235 mm based on the average of all cells of the PRISM 1971-2000 “normals” grid (PRISM Climate Group, Oregon State University, http://prism.oregonstate.edu, created 16 June 2006) contained within the catchment boundary (800m cell size). On average, greater than 85% of the annual precipitation occurs from October through April in “long-duration, low-to-moderate intensity frontal storms”[*Harr*, 1976]. Snow accumulation occurs occasionally, but is typically highly transient. Studies in the past have neglected snow as part of the precipitation record [*Harris*, 1977].

The bedrock underlying the research catchment is the Eocene-aged Tyee Formation of rhythmic-bedded layers of marine-derived greywacke sandstones and siltstones [*Snavely et al.*, 1964]. Boring logs from a series of wells installed throughout the Needle Branch catchment show that the shallow bedrock is weathered and fractured at least to a depth of 9 m below the ground’s surface (see *Hale and McDonnell* [this issue] for detailed core descriptions). Boring logs from two deep wells installed near the ridge of the upper catchment provide further evidence that the highly weathered and fractured zone extends to approximately 10 to 13 m dpeth (Figure 3.2). Well 1DP (DP=deep) was located at the inside of a road-cut; based on the exposed cut-section, we estimate that approximately 1.5 m of saprolite and 2 meters of fractured bedrock was removed from the location during road installation. Beds of fresh sandstone and siltstones begin at 11.5 m and continue to the installation depth (38.1 m) at Well-1DP. Water-bearing fractures were encountered between 30 and 35 m. Reconnaissance with a downhole camera indicated that these fractures were oriented parallel to the bedding plane and were the only water-bearing fractures observable within the borehole. Approximately 6 m of saprolite and 6 m of highly weathered and fractured rock overlay bedded layers of fresh sandstone and siltstone at Well-2DP (Figure 3.2). No water-bearing fractures were encountered during the drilling process at Well 2DP, nor were any observable with the downhole camera. The average bed thickness for both boreholes was in agreement with values reported by *Snavely et al.* [1964] for this formation (0.9 to 1.5 m). Cores were not obtained due to the drilling method employed (based on funding constraints), so detailed information on fracture density with depth is not available. However, a detailed description of a 35 m core drilled at the same landscape position and within the same geologic formation at the Mettman Ridge experimental hillslope was reported by *Anderson et al.* [1997b]. At Mettman Ridge, approximately 4 m of weathered rock was present directly below 0.5 m of colluvium. Fractured rock, approximately 4 m in thickness and having abundant oxidation features indicating water flow, was encountered below the weathered rock layer. Rock found below the fractured layer was characterized as fresh, although some fractures were present. Oxidized fractures were only present to 24 m of the 35 m borehole. Although this description matches well with the limited data exacted during drilling and camera reconnaissance at the two well installations in Needle Branch, the geology does not conform to this "neat" layer-cake structure across the entire catchment. Detailed topographic data obtained through airborne laser altimetry suggests several large areas of subdued topography that was likely created by deep-seated landslides—the presence of which has been documented to increase at this latitude within the Tyee Formation [*Roering et al.*, 2005].

Needle Branch hillslope soils are loams to gravelly loams (mesic Alic Hapludands and mesic Andic Humudepts) which average 1 m depth and are classified as well- to very well-drained [Corliss, 1973]. A ground-penetrating radar survey of two hillslopes in the upper Needle Branch catchment shows that although the soils are generally thin, soil depths can vary significantly over small distances (Figure 3.3). A more detailed description of the soils and their hydraulic properties is provided in *Hale and McDonnell* [this issue].

## METHODS

*Hydrometric Measurements*

Stream discharge was estimated at two locations in the Needle Branch catchment, NB-12 and NB-86 (number corresponds to catchment area in hectares; locations are marked on Figure 3.1), based on stage-discharge relationships developed at an H-type flume sealed onto a bedrock section of stream channel (NB-12) and a broad-crested, compound v-notch weir located at the end of a 3 m long, 3-sided (bottom and two sidewalls) concrete approach section (NB-86). Stream stage was measured at each control section with vented pressure transducers (Model PDCR 1830, GE Druck, Inc., Billerica, MA) and stored on a datalogger (Model CR-10X, Campbell Scientific, Inc., Logan, UT) at 10-min intervals. The period of record used in the analysis was October 2007 through September 2009 for NB-12 and October 2005 to September 2009 for NB-86 (NB-12 was not installed until October 2007). Precipitation inputs for NB-12 and NB-86 were taken as the areal average of a spatially distributed network of rain gauges located within the vicinity of Needle Branch (Figure 3.1).

Soil water content ($\theta_v$) was measured at a profile representing average hillslope soil conditions (Figure 3.1). We installed sensors (Model 10HS, Decagon Devices, Inc., Pullman, WA) at 0.15, 0.30, 0.65, and 0.85 m depths which were chosen to represent dominant textural classes present within the profile. The 0.15 and 0.30 m depths corresponded to upper and lower portions of a gravelly loam A-horizon. The 0.65 m depth was the mid-point in the gravelly, clay loam B-horizon and the 0.85 m depth represented the sandy clay loam BC horizon. Measurements were taken on 10 s intervals and 10-min mean values were stored on a datalogger (CR-10X. Profile water content was estimated on hourly timesteps by integrating $\theta_v$ for each sensor over its representative depth (0-0.225, 0.225-0.45, 0.45-0.75, and 0.75-1.00 m).

A network of groundwater wells was installed in the Needle Branch catchment. Two deep wells, Well-1DP and Well-2DP, are located approximately 150 m from each other near the ridge forming the northern boundary of the catchment (Figure 3.1). The total depth of Well-1DP is 38 m and Well-2DP is 60 m. Both DP wells were sealed into fresh rock with steel casing (at 15 m), backfilled with bentonite, and completed as 0.15 m diameter open boreholes. Three hillslope wells were installed into shallow bedrock in Needle Branch using the portable drilling system described by Gabrielli and McDonnell (2011); Well-3HS (HS=hillslope) and Well-8HS are located in the upper catchment (Figure 3.1) and Well-9HS is located in the lower catchment. The HS wells are all constructed of 25.4 mm inside-diameter (ID) polyvinylchloride (PVC). A 101.6 mm ID PVC outer casing was installed to the depth of the saprolite-bedrock transition. The inner casings were custom-made for each well so that the screened interval was sealed to a depth of 0.5 to 1 m below the saprolite-bedrock transition. The screen section was created using approximately 0.5 mm horizontal slots; the slotted section was covered with two layers of fiberglass mesh screen to prevent fine particles from entering the well. The annulus between the inner casing and the borehole wall was not large enough to allow backfilling with sand to an accurate, evenly packed depth, so the seal was created by positioning a rubber gasket on the outside of the inner casing such that it fit snuggly against the walls of the borehole at the desired sealing depth. Bentonite was backfilled on top of the gasket to the top of the outer casing to finalize the seal (approximately 0.1 m above the ground surface). Two floodplain (FP) wells, Well-10FP and Well-11FP, were installed in the lower Needle Branch catchment (Figure 3.1). The FP wells were constructed in the same manner as the HS wells and were sealed into the weathered bedrock underlying the surficial alluvial aquifer. Together, Well-9HS, Well-10FP, and Well-11FP lie along a transect lying orthogonal to the valley axis in lower Needle Branch. Water levels in all wells were measured on 10-min intervals with water level loggers (Model U20, Onset Computer Corporation, Inc., Pocasset, MA) and the records were corrected for barometric effects using an onsite barometric pressure measurement using the same sensor.

*Soil Water MTT Estimation*

Soil water MTT was estimated using the lumped-parameter convolution integral approach [M & Z, 1982]. The input characterization and modeling procedure employed is described in detail by *Hale and McDonnell* [this issue]. The tracer output for the transit time model was characterized by measuring the deuterium composition of soil water (0.5 m depth) collected from two suction lysimeters installed at mid-slope and toeslope positions in NB-12 near Welll-8HS (Figure 3.1). Samples were collected on daily to weekly intervals from June 2008 through June 2009. Following McGuire and McDonnell [2010], we used the dispersion flow model [*Maloszewski and Zuber*, 1982] to estimate the transit time distribution, $g(\tau)$, of soil water. The dispersion model is expressed as,

$$g(\tau) = \left(\frac{4\pi D_p \tau}{\tau_m}\right)^{-\frac{1}{2}} \tau^{-1} \exp\left[-\left(1-\frac{\tau}{\tau_m}\right)^2 \left(\frac{\tau_m}{4\pi D_p \tau}\right)\right] \tag{1}$$

where, $D_p$ is the dispersion parameter (1/Peclet number) and $\tau_m$ is the MTT.

*$^3$H-based Streamwater MTT Estimation*

The lumped-parameter convolution approach employed by *Hale and McDonnell* [this issue] to estimate deuterium-based MTTs was used to calculate $^3$H-based MTTs, except that a radioactive decay term was included in the calculation, so that,

$$C_{out}(t) = \int_0^\infty g(\tau)\, C_{in}(t-\tau)\, e^{-\lambda\tau} d\tau \tag{2}$$

where $C_{out}(t)$ is the $^3H$ concentration in streamflow at time t, $g(\tau)$ is the transit time distribution, $C_{in}$ is the precipitation $^3H$ concentration, and $e^{-\lambda\tau}$ is the radioactive decay term (with decay constant $\lambda = \ln(2/T_{1/2})$ and $T_{1/2}$=12.32 yr for $^3H$). We modeled $C_{out}$ based on weighted annual $C_{in}$ using the weighting function,

$$C_{in} = \sum_{i=1}^{12} C_i\, R_i / \sum R_i \tag{3}$$

where $C_i$ is the $^3H$ concentration in precipitation and $R_i$ is the recharge amount for month *i*. We estimated $R_i$ as the difference between monthly precipitation and monthly evapotranspiration. Monthly precipitation $^3H$ measured in Portland, OR [*Bob Michel*, unpublished report] was used for $C_i$. Where necessary, $C_i$ data gaps were filled using correlations with monthly measurements from Vienna, Austria (International Atomic Energy Agency, Global Network of Isotopes in Precipitation, http://www-naweb.iaea.org/napc/ih/IHS_resources_gnip.html). Stream $^3H$, used to define $C_{out}$, was sampled at NB-12 and NB-86 in June 2010 and analyzed at the GNS Science Water Dating Laboratory (Lower Hutt, New Zealand) using electrolytic enrichment and liquid scintillation counting [*Morgenstern and Taylor*, 2009]. The exponential-piston flow model was used to approximate $g(\tau)$. The two-parameter model is calculated as,

$$\begin{aligned} g(\tau) &= 0 && \text{for } \tau < \tau_m(1-f) \\ g(\tau) &= (f\tau_m)^{-1} \exp\left[-\left(\frac{\tau}{f\tau_m}\right) + \left(\frac{1}{f}\right) - 1\right] && \text{for } \tau \geq \tau_m(1-f) \end{aligned} \tag{4}$$

where *f* is the ratio of exponential volume to the total volume and $\tau_m$ is the MTT [*Maloszewski and Zuber*, 1982].

*Storage Estimation*

We leveraged the Mediterranean-type climate of the Pacific Northwest, USA, with distinct precipitation seasonality, to estimate dynamic storage volumes for the soil profile and the catchments, NB-12 and NB-86. Assuming a minimum residual storage volume at the end of the dry season, dynamic soil storage was calculated as the difference in maximum soil water content and soil water content at the end of the dry season. Similarly, dynamic catchment storage was estimated as the difference between maximum and minimum storage observed over the course of the dry-to-wet transition as calculated using the catchment water balance [e.g. *Sayama et al.*, 2011],

$$dV(t) = \sum_{t=1}^{T} P(t) - Q(t) - E(t) \tag{5}$$

where *dV(t)* is the cumulative change in storage from the beginning of the water year (October 1) to time *t*, *P(t)* is total precipitation, *Q(t)* is total discharge, and *E(t)* is evapotranspiration. We computed *E(t)* using the Penman-Monteith equation [*Monteith*, 1965] and meteorological variables measured onsite, with the exception of net radiation which was measured at a nearby Ameriflux site (Marys River Fir Site, approximately 28 km northeast of Needle Branch, http://public.ornl.gov/ameriflux/Site_Info/siteInfo.cfm?KEYID=us.oregon_fir.01).

We used stream discharge records and streamwater MTTs to approximate passive catchment storage, expressed mathematically as,

$$Passive\ Catchment\ Storage = MTTi \tag{6}$$

where *i* is the mean total annual discharge over the period MTT was estimated. Deuterium-based MTTs were calculated for NB-12 and NB-86 by convolving the tracer input signal (here, the deuterium composition in precipitation, reported as $\delta^2H$), weighted according to the gamma transit time distribution, with the tracer output measured in streamflow [e.g. *Maloszewski and Zuber*, 1982]. Details of field sampling, laboratory analysis, transit time modeling, and uncertainty analysis are provided in Hale and McDonnell (this issue). Tritium-based MTT estimation was described in the previous section (see *$^3$H-based MTT estimation*).

*Groundwater Age Determination*

We used tritium/helium-3 ($^3$H/$^3$He) dating to estimate the age of groundwater sampled at various depths and geomorphic positions within Needle Branch. The dating technique is based on the radioactive decay of $^3$H to $^3$He. When all other sources of $^3$He in groundwater can be accounted for (crustal and mantle sources), a mass balance between the $^3$H and the tritiogenic $^3$He in groundwater can be used to calculate the time elapsed since the sampled parcel of water became isolated from the atmosphere based on the known half-life of $^3$H [*Schlosser et al.*, 1988]. As a result of the low solubility and high diffusion coefficient of $^3$He, exchange of $^3$He with soil gas can occur readily within the vadose zone [*Solomon and Cook*, 2000]. Therefore, isolation from the atmosphere is typically not considered complete and, consequently, the $^3$H/$^3$He "clock" does not start, until the water has reached the saturated zone.

We collected water and dissolved gas samples from Well-1DP, Well2-DP, Well-8HS, and Well-11FP. In addition, we sampled a well located immediately adjacent to Well-11FP that was not previously described. This well, Well-16SF (SF=surficial), was installed to the soil-bedrock interface at 1 m below the ground surface and the screened interval (0.62 to 1 m) is located entirely within the soil profile. All wells were purged prior to sampling. The dissolved gas samples were collected with an advanced diffusion sampler [*Gardner and Solomon*, 2009] which was deployed for approximately 72 hours following purging. Samples were analyzed for $^3$H and dissolved gas concentrations at the University of Utah Noble Gas Laboratory (Salt Lake City, UT).

**RESULTS**

*Hydrodynamic and deuterium characterization*

Precipitation, specific discharge, precipitation and streamwater $\delta^2H$, and groundwater levels for the period of 01 October 2009 to 30 May 2010 are shown in Figure 4.4. This period covers the duration of the wet season for our Pacific Northwest catchments, including the transition from dry to wet conditions (hereafter referred to as the "wet-up period"). The hydrograph traces in Figure 4.4a show that the streams responded rapidly to precipitation inputs after the initial wet-up period (a quantitative analysis of the hydrodynamics of these catchments is provided in *Hale and McDonnell* [this issue]. Specific discharge was higher at the smaller NB-12 catchment during most runoff events, however the mean difference in specific discharge between the catchments during the period shown was only 0.04 mm $hr^{-1}$.

Despite the highly-responsive rainfall-runoff dynamics of these catchments, the $\delta^2H$ signal in streamwater was substantially damped relative to the variability of $\delta^2H$ measured in precipitation (Figure 4.4b). The standard deviation of streamwater $\delta^2H$ was 1.8 ‰ for NB-12 and 2.2 ‰ for NB-86 over the entire period of record while the standard deviation of precipitation $\delta^2H$ was 15.1 ‰ (Table 4.1). The amount-weighted mean $\delta^2H$ of precipitation was -50.6 ‰; the mean streamwater $\delta^2H$ compositions were -49.6 ‰ and -50.1 ‰ for NB-12 and NB-86, respectively. For comparison purposes, the standard deviation of $\delta^2H$ for the period shown in Figure 4.4b was 1.3 ‰ for NB-12, 1.9 ‰ for NB-86, and 16.4 ‰ for precipitation. The amount-weighted mean $\delta^2H$ of precipitation was -57.3 ‰ during this period while the mean streamwater $\delta^2H$ compositions were -48.4 ‰ and -48.0 ‰ for NB-12 and NB-86, respectively.

Each of our two deep wells displayed unique hydrodynamic characteristics during the 2010 wet season (Figure 4.4c). Although water levels in both wells increased in elevation by approximately the same absolute value (roughly 4 m), Well-1DP was distinctly more sensitive to precipitation events than Well-2DP. At Well-2DP, the water table increased gradually over the course of the wet season whereas, at Well-1DP, many short-term fluctuations occurred as the water level increased. Hillslope wells Well-3HS (Figure 4.4d) and Well-9HS (Figure 4.4e) were both non-responsive during the early wet-up period but began to rapidly rise to a new base level after 59 mm and 222 mm, respectively, of precipitation accumulated. Once activated, these wells responded rapidly to precipitation throughout the remainder of the wet season. The other hillslope well, Well-8HS, did not display a threshold-type behavior prior to responding to precipitation. Instead, its water level closely mimicked the dynamics of the stream for the entire wet season, including the wet-up period (Figure 4.4d). Both floodplain wells, Well-10FP and Well-11FP (sealed into the bedrock below the alluvial sediments), were responsive to precipitation during the wet-up period and throughout the wet season but the range of fluctuations were less than 0.5 m and 0.2 m, respectively, over the entire period.

We did not sample the wells during the 2010 wet season since the act of taking a sample contaminated the water level record and would obscure interpretation. We did however, collect samples for isotopic analysis prior to this period. Table 4.1 provides the mean $\delta^2H$ values and standard deviations for the sampled groundwater. The mean values for all groundwaters were within 2 ‰ of the mean precipitation and streamwater values (analytical error=1 ‰). Similar to streamwater, the groundwater $\delta^2H$ values did not vary significantly as indicated by the standard deviations of all mean values being less than 2 ‰.

*Soil Water MTT*

Soil water MTT modeling results are reported in Table 3.2. The maximum likelihood MTT estimate for soil water collected at the midslope lysimeter was 132 days. The toeslope MTT was approximately half of the midslope value (63 days). At both locations, the model performed reasonably well (Nash-Sutcliffe efficiencies of 0.57 and 0.79, respectively). The tenth and ninetieth percentile MTT estimate from the log-likelihood uncertainty estimation procedure (see Hale and McDonnell [this issue] for a detailed description of the method) produced an uncertainty range of 38 days for the midslope lysimeter (96 to 134 days) and 10 days for the toeslope lysimeters (59 to 69 days).

### *$^3$H-based Streamwater MTT*

Mean transit times were estimated from single streamwater $^3$H values at NB-12 and NB-86 by iteratively adjusting the two parameters of exponential-piston flow model, $f$ and $\tau_m$ (where $\tau_m$ =MTT), until the parameter set that matched the measured $^3$H concentration of the sample to the model output, $C_{out}$, was identified The *a priori* parameter distributions were $0<=f<=1$ and $0< \tau_m<=100$ yrs. For both NB-12 and NB-86, the best fit was found for $f$=0.7 (Table 3.3) although $f$ values from 0.4 to 0.8 all provided similar MTT estimates. The best-fit $^3$H-based MTTs were 6.8 yrs and 7.7 yrs for NB-12 and NB-86, respectively; but because of the uncertainty associated with this approach, we used the lower and upper bounds of the MTT estimates in our storage analysis (reported in Table 3.3).

Potential uncertainty sources in our MTT estimates are sample measurement error, error in our flow model, and/or error in the input function. The sample measurement error was ± 0.04 TU, which is associated with an MTT error of ± 0.4 yrs. If we assume the error in the flow model estimation was ± 10% (i.e. $f$=70 ± 10 %), the associated MTT error is ± 0.5 yrs. Errors in the input function result from sample measurement error for precipitation, correlation errors with the Vienna precipitation record used to fill data gaps in the Portland, Oregon precipitation time series, and errors estimating the recharge weighting function, together estimated to total ± 0.2 TU which corresponds to a MTT error of ± 1.6 yrs. Assuming these three error sources are independent, the overall MTT error was ± 1.7 yrs. This value was applicable to both NB-12 and NB-86 since both samples were subject to the same error sources.

*Dynamic and Passive Storage*

Soil water content from a representative soil profile was used to estimate the dynamic storage available within the soil across the catchments. Figure 3.5 shows stream discharge as a function of soil water content during the course of the wet season, including the wet-up period (see inset for the source data time series). The dynamic soil storage for this measurement period was 120 mm. Maximum soil storage for this period was 235 mm. A threshold behavior was evident in the stream discharge versus soil water relationship, whereby stream discharge did not commence until soil moisture storage reached 180 to 200 mm. This threshold value represents 77 to 85 % of the maximum soil storage for our measurement period (235 mm).

Cumulative change in storage volume, *dV*, since the beginning of each water year of record was estimated for NB-12 and NB-86. Figure 3.6 shows stream discharge as a function of *dV* for NB-12 (a) and NB-86 (b). Runoff generation occurred across a range of *dV* values at both catchments, except approximately the lowest 100 mm interval. This “threshold” value aligns well with our estimate of 120 mm of soil profile dynamic storage. Catchment dynamic storage, taken as the absolute difference between the maximum *dV* and minimum *dV* measured during the period of record, was 485 mm for NB-12 and 501 mm for NB-86.

We used MTT estimations derived from both stable isotope ($\delta^2H$) and $^3H$ values of NB-12 and NB-86 streamwater to estimate passive storage within each catchment (Figure 3.7; dynamic storage estimates are also plotted for comparison). As a result of the uncertainty associated with the MTT estimations, we used a conservative approach to calculate the potential range of passive storage. The lower bound of the MTT estimate was multiplied by the lower bound of the mean of the total annual discharge values (taken as mean total annual discharge minus 10 %) for the period of record over which the $\delta^2H$ samples were collected for MTT estimation to estimate the lower limit of passive storage. Likewise, the upper bounds of the MTT estimate and mean of the total annual discharge (taken as mean total annual discharge plus 10 %) were used to define the upper limit of passive storage. The resulting ranges of passive storage for NB-12 were 5900 to 15600 mm based on the $\delta^2H$ MTT estimate and 8500 to 13300 mm based on $^3H$. At NB-86, passive storage is estimated to range from 4900 to 8400 mm ($\delta^2H$) and 8400 to 13100 mm ($^3H$). The close agreement of both MTT-based storage estimates ($\delta^2H$- and $^3H$-based) for each catchment is evident in Figure 3.7. Passive storage for both NB-12 and NB-86 is more than an order of magnitude larger than dynamic storage.

*Groundwater Ages*

We used $^3H/^3He$ dating to estimate groundwater ages at five wells within the NB catchment (Figure 3.8). Groundwater was youngest (2.8 ±2.3 yrs) in Well-8HS, which is located approximately 10 m (horizontal distance) from the stream and integrates fractures occurring in the bedrock between 2.5 and 5.7 m below the ground surface. The groundwater in Well-1DP and Well-2DP was approximately 2 years older than the shallow bedrock water in Well-8HS (5.0 ±1.8 yrs and 5.8 yr[2], respectively). Lower in the catchment, the bedrock groundwater below the alluvial floodplain was greater than 55 yrs-old (Well-11FP; an exact age was not possible given its extremely low, pre-bomb era $^3H$ concentration). Age determination for Well-16SF, the surficial aquifer immediately overlying the Well-11FP unit, was not possible but the sample appeared to be a mixture of modern and old water based on the $^3H$ concentration (intermediate to the old groundwater below it and the younger, shallow

[2] The uncertainty calculation was not possible.

bedrock groundwater in Well-8HS) and because the ratio of $^3H/^3He$ in the sample was the same as the $^3H/^3He$ ratio of the atmosphere. The interpretation of mixed water in the surficial aquifer is supported by hydrometric measurements that showed an upward, vertical hydraulic gradient for the underlying bedrock groundwater; this indicated that the older bedrock groundwater was upwelling based on the position of the piezometric surfaces (see depiction in Figure 3.8).

## DISCUSSION

*Subsurface storage characterization: the bottom-up approach*

We used a multi-pronged approach to investigate the subsurface storage controls on streamwater MTT and MTT scaling in the permeable sandstone geology catchments studied by *Hale and McDonnell* [this issue]. By combining direct interrogation of the subsurface through borehole characterization and groundwater monitoring with tracer-based dating techniques, we identified discrete storage zones that influence streamwater MTT. Although a continuum system, our bottom-up approach noted five such distinct zones with distinct storage controls: (1) soil storage, i.e. the pore space in the soil and subsoil available to hold water. This soil water retention can be short as evidenced by our lysimeter-based MTT estimates for soil water on the order of 60 to 130 days, (2) shallow bedrock storage, including the saprolite and the highly, fractured upper layer bedrock; based on our borehole data, the thickness of this storage zone likely ranges from a couple of meters to over ten meters, (3) the deep bedrock zone; this  consists of storage available in rock matrix and fractures in bedrock below the shallow bedrock zone; the lower boundary of this storage zone is unknown, (4) the surficial alluvial zone; this is comprised of the soil and alluvial and colluvial sediments lying above the bedrock in the widened valley-bottom with alluvial plain geomorphology. The soil texture in the alluvial plain grades from a loam at the surface to a silty clay at depth, inducing at least partial saturation of the profile for most of the year and creating an aquitard between the surficial alluvial zone and the last identifiable storage zone, (5) the sub-alluvial zone or storage held in the weathered and fractured bedrock lying immediately below the surficial alluvial zone; the depth of this zone is unknown, but it is at least 5 m thick based on our borehole data.

Storage volume in the soil zone is small relative to estimates of passive catchment storage. Maximum soil storage during our measurement period was 235 mm, compared to passive storage estimates for the entire catchment on the order of 10,000 mm. This ratio of soil water storage to catchment storage is very different to the more traditional poorly-permeable headwater research catchment, where soil water storage is the largest of all the mobile storages in the system (see Sayama and McDonnell [2009] for review). Our maximum soil storage estimate is based on a single wet-season and is therefore best taken as an approximate maximum storage. The theoretical maximum for soil storage is 700 mm based on a 1 m profile with uniform porosity of 70 %. Notwithstanding uncertainty in these calculations, both are extremely small relative to total catchment passive storage. The threshold behavior between soil storage and stream discharge indicates that, in most cases, storage deficits in the soil profile must be satisfied before runoff is generated (a common finding in headwater systems [*Western et al.*, 2002]). Water storage within the soil is highly transient relative to the groundwater storage zones, with MTTs estimated between 60 and 130 days. Water flow direction in the same soil type at the Mettman Ridge site was exclusively vertical under all wetness conditions [*Torres et al.*, 1998]. We therefore view the soil profile as a temporary storage zone that regulates vertical recharge to the shallow bedrock zone immediately below.

Groundwater dynamics in the shallow bedrock zone varied among our three wells, indicating heterogeneity in the storage properties of this zone. Such heterogeneity is expected in highly fractured rock aquifers [*Freeze and Cherry*, 1979].

Two of the hillslope wells (HS series) did not respond immediately to precipitation during the wetting-up period, but after a presumable storage deficit was replenished, the groundwater increased rapidly to a new base water level and became responsive to further precipitation inputs (Figure 3.4). This delayed, threshold-like response was also observed in shallow bedrock groundwater at the Mettman Ridge hillslope [*Anderson et al.*, 1997b; *Montgomery et al.*, 2002] and fits the *Sidle et al.* [2001] hydrogeomorphic conceptual model whereby different geomorphic units become hydrologically-active with increasing antecedent wetness. Well-8HS behaved differently than the other two hillslope wells, closely matching the dynamics of streamflow independent of antecedent conditions. In combination with the observed groundwater dynamics, extended pumping of Well-8HS, at rates of up to 2 liters per minute during the late-summer (i.e. driest antecedent conditions), provided convincing evidence that this borehole intersects perennially water-bearing fractures that are tightly connected to the stream. The groundwater in this well was younger than the MTT of the streamwater approximately 100 m downstream at NB-12 (3 yrs versus 5 yrs), indicating that other (and older) storage components contribute to streamflow at the small, 12-ha catchment scale.

Groundwater dynamics in our two deep bedrock wells displayed different characteristic behavior, similar to the heterogeneity observed in the shallow bedrock wells. Water levels in both deep wells increased over the course of the wet-season (Figure 3.4). However, the responsive nature of Well-1DP suggests that the water-bearing fractures mapped at 30 to 35 m depth in this borehole are indeed hydrologically relevant on the timescale of individual storm events. While the responses are likely expressions of pressure wave propagation through a hydraulically-primed fracture network [*Rasmussen*, 2001], the timescale of the response indicates an intimate connection between the surface and deeper storage zones under wet antecedent conditions. The groundwater dynamics in Well-2DP are much more subdued, following a gradual increase over the course of the wet season with no response to individual precipitation events. Despite the dissimilarities in water level dynamics on short timescales, the age of the groundwater in these two wells is similar; 5 and 6 yrs for Well-1DP and Well-2DP, respectively. The groundwater elevation in Well-2DP is, on average, 10 m higher than the water surface in Well-1DP 150 m away (the ground surface elevation is 20 m higher at Well-2DP). With a depth differential of only 10 m at 150 m distance and similar ages, we expect that we are measuring the same groundwater storage component. It is also likely that stored water in the deep bedrock zone contributes to streamflow, as inferred from the mixing with the younger, shallow bedrock water necessary to achieve the streamwater MTT measured at NB-12.

Groundwater in the sub-alluvial zone is the first to exhibit a response to precipitation inputs early in the wetting-up phase, reacting even before the stream begins to respond to precipitation. We believe this storage zone is semi-confined as a result of the silty clay aquitard created at the base of the soil profile in the alluvial plain. The piezometric surface elevation confirms some level of confinement in this unit as it is always above the grade of the free water surface in the surficial alluvial zone (not shown), and becomes artesian during larger storm events. This indicates an upward flow gradient into the surficial alluvial zone that becomes stronger during storm events. The $^3$H concentration, as well as evidence from dissolved silica analysis and specific conductivity measurements made at the time of $^3$H sampling, confirm that this is a distinctly different pool of water than that present in the surficial alluvial zone above it. The $^3$H concentration was 0.30 tritium units (TU) in the sub-alluvial zone compared to 1.46 TU in the surficial zone; the silica concentration was 10.2 mg $l^{-1}$ and specific conductivity was 264 μS $cm^{-1}$ in the sub-alluvial groundwater (Well-11FP) compared to 4.46 mg $l^{-1}$ and 34 μS $cm^{-1}$ in the surficial alluvial groundwater (Well-16SF). This distinction was not evident based on the $\delta^2$H and $\delta^{18}$O values of the two water pools (see similar values in Table 3.1). At greater than 55 yrs-old, the water in the sub-alluvial zone is considerably older than any of the other storage zones we identified. This finding links directly to the issue of stable isotope-based MTT truncation of reported by *Stewart et al.* [2010]. If additional tracers were not employed, we would not have been able to discriminate this zone from the surficial alluvial zone above.

Groundwater dynamics for the surficial alluvial zone are not available for the measurement period shown in Figure 3.4 as a result of instrument failures. However, monitoring from previous years shows that the water table dynamics closely follow the stream (not shown). Groundwater age dating was inconclusive but the ratio of the sample $^3$H concentration to the $^3$H concentration of the atmosphere showed strong evidence of mixing of old sub-alluvial and more modern hillslope water. Similar mixing of old and young water was reported by *Solomon et al.* [2010] at forested catchments in Costa Rica. The surficial alluvial zone therefore acts as a mixing tank, combining water from two distinctly different storage reservoirs prior to discharging it to the stream.

*Radically different plumbing hidden by similar catchment forms*

Our results suggest that the bedrock permeability in the sedimentary catchments of the Oregon Coast Range creates a groundwater flow system where streamflow in the smaller catchments is primarily fed by local groundwater flow in the shallow and deep bedrock zones. As catchment scale increases, contributions from the surficial alluvial and sub-alluvial zones become important. The sub-alluvial zone is conceptualized as a landscape-scale storage reservoir that is regionally recharged through the deep fracture network. The surficial alluvial zone integrates young and local shallow bedrock groundwater moving laterally off of the adjacent hillslopes with the older, more regional groundwater upwelling from below the valley-bottom. The proportions of young hillslope groundwater and old valley-bottom groundwater combining in this critical mixing zone sets the MTT of the groundwater discharging to the stream. Streamwater MTT is then set by the mixing of streamwater already in the channel with local groundwater inflow.

The MTT scaling observed in our permeable bedrock catchments, whereby streamwater MTT increases with catchment area, appears largely controlled by the mixing that occurs in the surficial alluvial zone. The multitude of small, 1$^{st}$- and 2$^{nd}$-order catchments, present as a result of the highly dissected landscape, do not have a significant alluvial plain themselves and therefore the MTTs at small catchment scales are set by the combination of groundwater discharge from the shallow and deep bedrock zones, as observed in NB-12. However, at larger scales, where the surficial and sub-alluvial zones become relevant, MTTs increase as increasing proportions of old, sub-alluvial water contributes to streamflow. Relative contributions from the sub-alluvial zone likely increase in a down-valley direction, and hence, with increasing catchment area as a function of the increasing size of the sub-alluvial zone resulting from increasing accumulated recharge area. Additionally, Personious [1995] showed that stream channel incision into the bedrock proceeds in a downstream direction in Oregon Coast Range streams (Drift Creek was included in his study). As the channel cuts closer and/or into the underlying sub-alluvial zone, contributions from this zone would be expected to increase based on such a geomorphological rationale; leading to concomitant increases in streamwater MTTs with increasing catchment area.

The multi-component storage system we find in the permeable sedimentary bedrock at our Oregon Coast Range catchments contrasts with the much simpler storage system in the volcanic HJA catchments as described in *Hale and McDonnell* [this issue]. The tight bedrock geology restricts storage to the soil profile on the HJA hillslopes. Recent borehole investigations by *Gabrielli and McDonnell* [2011] indicate some potential for storage in shallow fractures in the volcanic bedrock of the HJA. However, there is little vertical connectivity below this thin, fractured zone and effectively impermeable rock below, so this storage zone likely has less of an effect on setting streamwater transit times and more influence on controlling the precipitation threshold for stream response as observed by *Graham and McDonnell* [2010]. Indeed, detailed, bottom-up investigations at the HJA [*Harr*, 1977; *McGuire and McDonnell*, 2010] show the strong permeability contrast between the soil and bedrock results in precipitation-induced development of a thin, transient zone of saturation at the soil-bedrock interface which moves laterally downslope as gravity-driven flow. These gravity- and topographic-driven runoff processes relate directly to the relatively short MTTs and MTT relationships with flowpath length and gradient observed at the HJA catchments [*McGuire et al.*, 2005]. Although surficial alluvial zones exist in catchments at the HJA that have valley-bottoms morphology [*Swanson and James*, 1975], the MTTs presented in *McGuire et al.* [2005] suggest that storage within the surficial alluvial zone does not significantly affect catchment transit time distributions. This is shown clearly by the similarity in streamwater MTTs at WS10 (1.2 yrs), which has no riparian storage, and WS03 (1.3 yrs), where *Wondzell* [2006] documented storage within the alluvial and colluvial sediments that fill the valley-bottom. The sharp permeability contrast at the soil-bedrock interface created by the tight volcanic geology at the HJA results in a system where storage is relegated to the soil profile and runoff processes and streamwater MTTs are driven by topography. The contrasts in landscape-scale anisotropy may in fact be the simplest way to differentiate our findings in these two broad classes of catchments: low landscape anisotropy in the permeable Oregon Coast range, where no ‘throttle’ for lateral subsurface flow exists [*Bonell and Bruijnzeel*, 2004], to high landscape-scale anisotropy at HJA whereby the throttle exists and is high throughout the wet season. Indeed the latter is the most widely reported class of headwater system (see similar features at the Maimai water in New Zealand; Fudoji watershed in Japan; Plastic Lake watershed in Canada).

*Fenicia et al.* [2010] showed that MTT estimates based on lumped-parameter MTT models were quite sensitive to the mixing assumptions implicit to the theory. Although theoretical progress is being made to account for the mixing of multiple storage (or landscape) units with unique transit time distributions [*Botter et al.*, 2010; *Rinaldo et al.*, 2011], little work has been done to verify the new theory at a process-level in the field. Our findings document the importance of these mixing processes for controlling MTT and MTT scaling behavior and represent a starting point for more field-based work to test the new advances in theory.

*Dynamic versus passive storage volumes*

Passive storage at NB-12 and NB-86 was an order of magnitude larger than the water balance-derived dynamic storage estimations (soil and catchments). Neither passive nor dynamic storage showed a dependence on catchment scale. Many factors contribute to uncertainty in the dynamic storage estimates, such as errors in precipitation, stream discharge, and evapotranspiration estimates. Likewise, heterogeneity in soil depth and porosity contribute to uncertainty in estimating dynamic storage within the soil. Even accounting for considerable uncertainty in our estimations (we conservatively estimated uncertainty as ±50 % of the calculated dynamic storage estimate), passive storage is still significantly larger than dynamic storage ($10^3$-$10^4$ versus $10^2$, respectively) at both NB-12 and NB-86 (Figure 3.7). While this difference seems extreme, *Birkel et al.* [2011a] also observed order of magnitude differences in dynamic and passive storage at two catchments in the Scottish Highlands. They attributed the relatively small dynamic storage to a combination of thin hillslope soils overlying impermeable bedrock and highly responsive valley-bottom soils that overly poorly permeable glacial till. The relatively larger passive storage is a result of the overall volume of saturated soil present in the wide, glacially-carved valley. At our Oregon Coast Range catchments the dynamic storage measured in the soil profile (approximately 120 mm) in combination with the amount of precipitation necessary to activate shallow bedrock groundwater in two of our shallow bedrock wells (60 and 220 mm) is in close agreement with the catchment-scale dynamic storage estimates (485 and 500 mm for NB-12 and NB-86). Since dynamic storage is relatively small and can essentially be accounted for by storage in the shallow subsurface, the considerable passive storage volume is likely achieved by a continually full deeper fracture system.

*Sayama et al.* [2011] linked the dynamic storage volume of various-sized catchments in coastal northern California to the mean catchment slope angle, suggesting that steeper catchments were capable of storing greater volumes of water in the hydrologically-active bedrock. At our site, NB-12 is significantly steeper than NB-86 (median slope of 51 versus 34 %, respectively), however the dynamic storage estimates are relatively similar (within 15 mm). Although our sample size is small, our results do not support the *Sayama et al.* [2011] hypothesis. The lower slope angles in NB-86 are linked to a past history of large deep-seated landslides that have not occurred in NB-12. It is possible that, during the course of the landslide, any loss of storage associated with the size of the overall control volume (based on high slope angles) was offset by an increase in the volume of void space created during fracturing and crumbling of bedrock during the slide. Given the importance of subsurface storage outlined in this study, the role of deep-seated landslides in controlling catchment storage is an interesting, and open question in the hydrology of our system.

## CONCLUSIONS

In this paper, we showed the importance of bedrock permeability in controlling streamwater MTTs and MTT scaling relationships. We found that the permeable sedimentary bedrock in our Oregon Coast Range catchments led to the development of distinct zones of storage. Our hydrometric, MTT, and groundwater dating analyses showed that streamwater MTTs were controlled by a mixture of water contributions from the individual zones within the permeable bedrock. We hypothesize that the relative contributions from each storage component changes with catchment area and leads to the MTT scaling behavior that we observed. Without our bottom-up process-based approach—employing multiple tracers and dating methods—this mechanistic understanding of how MTTs are controlled and scaled based on catchment storage characteristics would not have been possible. Our work suggests that permeability contrasts in the subsurface represent perhaps the most basic control on how water is stored within the subsurface and therefore is perhaps one of the best direct predictors of streamwater MTT (i.e. more than previously-derived morphometric-based predictors).

Efforts are underway within the hydrological research community to derive a catchment classification system [*McDonnell and Woods*, 2004; *Wagener et al.*, 2007] and inclusion of catchment function (collection, storage and discharge) into any classification scheme is important. Our work shows that, first and foremost, such a functional representation requires thorough understanding of the underlying geology and, more specifically, how the geologic substrate and soils create permeability contrasts within the subsurface. Furthermore, our work suggests that indices of catchment form and rainfall-runoff response are incomplete descriptors of catchment function as they did not fully account for catchment storage. Our overarching message is that, just like beauty, catchment function is more than skin-deep and efforts to classify or group catchments should begin to include such metrics as shown to be important for the ranges of MTT and scaling relations as evidenced in this work and in *Hale and McDonnell* [this issue].

## REFERENCES

Ali, G. A., C. L'Heureux, A. G. Roy, M.-C. Turmel, and F. Courchesne (2011), Linking spatial patterns of perched groundwater storage and stormflow generation processes in a headwater forested catchment, *Hydrol. Processes*, n/a-n/a.

Anderson, S. P., W. E. Dietrich, R. Torres, D. R. Montgomery, and K. Loague (1997a), Concentration-discharge relationships in runoff from a steep, unchanneled catchment., *Water Resour. Res.*, *33*(1), 211-225.

Anderson, S. P., W. E. Dietrich, D. R. Montgomery, R. Torres, M. E. Conrad, and K. Loague (1997b), Subsurface flow paths in a steep, unchanneled catchment., *Water Resour. Res.*, *33*(12), 2637-2653.

Asano, Y., T. Uchida, and N. Ohte (2002), Residence times and flow paths of water in steep unchannelled catchments, Tanakami, Japan, *J. Hydrol.*, *261*(1-4), 173-192.

Birkel, C., C. Soulsby, and D. Tetzlaff (2011a), Modelling catchment-scale water storage dynamics: reconciling dynamic storage with tracer-inferred passive storage, *Hydrol. Processes*, n/a-n/a.

Birkel, C., D. Tetzlaff, S. M. Dunn, and C. Soulsby (2011b), Using time domain and geographic source tracers to conceptualize streamflow generation processes in lumped rainfall-runoff models, *Water Resour. Res.*, *47*(2), W02515.

Bonell, M. (1993), Progress in the understanding of runoff generation dynamics in forests, *J. Hydrol.*, *150*, 217-275.

Bonell, M., and L. A. Bruijnzeel (Eds.) (2004), *Forests, Water and People in the Humid Tropics: Past, Present and Future Hydrological Research for Integrated Land and Water Management*, Cambridge University Press, Cambridge.

Botter, G., E. Bertuzzo, and A. Rinaldo (2010), Transport in the hydrologic response: Travel time distributions, soil moisture dynamics, and the old water paradox, *Water Resour. Res.*, *46*.

Botter, G., A. Porporato, I. Rodriguez-Iturbe, and A. Rinaldo (2009), Nonlinear storage-discharge relations and catchment streamflow regimes, *Water Resour. Res.*, *45*.

Broxton, P. D., P. A. Troch, and S. W. Lyon (2009), On the role of aspect to quantify water transit times in small mountainous catchments, *Water Resour. Res.*, *45*.

Brutsaert, W., and J. L. Nieber (1977), REGIONALIZED DROUGHT FLOW HYDROGRAPHS FROM A MATURE GLACIATED PLATEAU, *Water Resour. Res.*, *13*(3), 637-644.

Fenicia, F., S. Wrede, D. Kavetski, L. Pfister, L. Hoffmann, H. H. G. Savenije, and J. J. McDonnell (2010), Assessing the impact of mixing assumptions on the estimation of streamwater mean residence time, *Hydrol. Processes*, *24*(12), 1730-1741.

Freeze, R. A., and J. A. Cherry (1979), *Groundwater*, 1st ed., 604 pp., Prentice-Hall, Englewood Cliffs, NJ.

Gabrielli, C. P., and J. J. McDonnell (2011), An inexpensive and portable drill rig for bedrock groundwater studies in headwater catchments, *Hydrol. Processes*, n/a-n/a.

Gardner, P., and D. K. Solomon (2009), An advanced passive diffusion sampler for the determination of dissolved gas concentrations, *Water Resour. Res.*, *45*.

Graham, C. B., and J. J. McDonnell (2010), Hillslope threshold response to rainfall: (2) Development and use of a macroscale model, *J. Hydrol.*, *393*(1-2), 77-93.

Hale, V. C., and J. J. McDonnell (this issue), Similar catchment forms and rainfall-runoff responses hide radically different plumbing: (1) A top-down analysis of mean transit time ranges and scaling relations, *to be submitted to Water Resources Research*.

Haria, A. H., and P. Shand (2004), Evidence for deep subsurface flow routing in a forested upland Wales: implications for contaminant transport and stream flow generation, *Hydrol. Earth Syst. Sci.*, *8*(2), 334-344.

Harr, R. D. (1976), Hydrology of small forest stream in western Oregon, General Technical Report PNW-55*Rep.*, 15 pp, Pacific Northwest Forest and Range Experiment Station, USDA Forest Service, Portland, OR.

Harr, R. D. (1977), Water flux in soil and subsoil on a steep forested slope, *J. Hydrol.*, *33*, 37-58.

Harris, D. D. (1977), Hydrologic changes after logging in two small Oregon coastal watersheds, *United States Geological Survey*, *Water-Supply Paper 2037*.

Hewlett, J. D., and A. R. Hibbert (1967), Factors affecting the response of small watersheds to precipitation in humid areas, in *Forest Hydrology.*, edited by W. E. Sopper and H. W. Lull, pp. 275-291, Pergamon Press, New York.

Hrachowitz, M., C. Soulsby, D. Tetzlaff, J. J. C. Dawson, and I. A. Malcolm (2009), Regionalization of transit time estimates in montane catchments by integrating landscape controls, *Water Resour. Res.*, *45*.

Katsuyama, M., M. Tani, and S. Nishimoto (2010), Connection between streamwater mean residence time and bedrock groundwater recharge/discharge dynamics in weathered granite catchments, *Hydrol. Processes*, *24*(16), 2287-2299.

Kirchner, J. W. (2009), Catchments as simple dynamical systems: Catchment characterization, rainfall-runoff modeling, and doing hydrology backward, *Water Resour. Res.*, *45*.

Kosugi, K. i., S. y. Katsura, M. Katsuyama, and M. Takahisa (2006), Water flow processes in weathered granitic bedrock and their effects on runoff generation in a small headwater catchment, *Water Resour. Res.*, *42*.

Kosugi, K. i., M. Fujimoto, S. y. Katsura, H. Kato, Y. Sando, and T. Mizuyama (2011), Localized bedrock aquifer distribution explains discharge from a headwater catchment, *Water Resour. Res.*, *47*(7), W07530.

Maloszewski, P., and A. Zuber (1982), Determining the turnover time of groundwater systems with the aid of environmental tracers. 1. models and their applicability, *J. Hydrol.*, *57*, 207-231.

Manning, A. H., and J. S. Caine (2007), Groundwater noble gas, age, and temperature signatures in an Alpine watershed: Valuable tools in conceptual model development, *Water Resour. Res.*, *43*(4).

McDonnell, J. J., and R. Woods (2004), On the need for catchment classification, *J. Hydrol.*, *299*(1-2), 2-3.

McDonnell, J. J., et al. (2010), How old is streamwater? Open questions in catchment transit time conceptualization, modelling and analysis, *Hydrol. Processes*, *24*(12), 1745-1754.

McGuire, K. J., and J. J. McDonnell (2006), A review and evaluation of catchment transit time modeling, *J. Hydrol.*, *330*(3-4), 543-563.

McGuire, K. J., and J. J. McDonnell (2010), Hydrological connectivity of hillslopes and streams: Characteristic time scales and nonlinearities, *Water Resour. Res.*, *46*.

McGuire, K. J., J. J. McDonnell, M. Weiler, C. Kendall, B. L. McGlynn, J. M. Welker, and J. Seibert (2005), The role of topography on catchment-scale water residence time, *Water Resour. Res.*, *41*(5), -.

McNamara, J. P., D. Tetzlaff, K. Bishop, C. Soulsby, M. Seyfried, N. E. Peters, B. T. Aulenbach, and R. Hooper (2011), Storage as a metric of catchment comparison, *Hydrol. Processes*, n/a-n/a.

Millares, A., M. J. Polo, and M. A. Losada (2009), The hydrological response of baseflow in fractured mountain areas, *Hydrol. Earth Syst. Sci.*, *13*(7), 1261-1271.

Monteith, D. T. (1965), Evaporation and the environment, *Symposium of the Society for Experimental Biology*, *19*, 205-234.

Montgomery, D. R., W. E. Dietrich, and J. T. Heffner (2002), Piezometric response in shallow bedrock at CB1: Implications for runoff generation and landsliding, *Water Resour. Res.*, *38*(12), doi:10.1029/2002WR001429.

Morgenstern, U. (2007), Lake Taupo Streams: Water age distribution, fraction of landuse impacted water, and future nitrogen load*Rep.*, 21 pp, Environment Waikato, Hamilton, New Zealand.

Morgenstern, U., and C. B. Taylor (2009), Ultra low-level tritium measurement using electrolytic enrichment and LSC, *Isotopes in Environmental and Health Studies*, *45*(2), 96-117.

Morgenstern, U., R. R. Reeves, C. J. Daughney, S. Cameron, and D. Gordon (2005), Groundwater age and chemistry, and future nutrient load for selected Rotorua lakes catchments*Rep.*, 73 pp.

Niemi, A. J. (1978), Residence time distribution of variable flow processes, *International Journal of Applied Radiation and Isotopes*, *28*, 855-860.

Oswald, C. J., M. C. Richardson, and B. A. Branfireun (2011), Water storage dynamics and runoff response of a boreal Shield headwater catchment, *Hydrol. Processes*, *25*(19), 3042-3060.

Rasmussen, T. C. (2001), Pressure wave vs. tracer velocities through unsaturated fractured rock, *Flow and transport through unsaturated fractured rock; an overview*, *42*, 45-52.

Reager, J. T., and J. S. Famiglietti (2009), Global terrestrial water storage capacity and flood potential using GRACE, *Geophys. Res. Lett.*, *36*.

Rinaldo, A., K. J. Beven, E. Bertuzzo, L. Nicotina, J. Davies, A. Fiori, D. Russo, and G. Botter (2011), Catchment travel time distributions and water flow in soils, *Water Resour. Res.*, *47*.

Robinson, D. A., et al. (2008), Advancing process-based watershed hydrological research using near-surface geophysics: a vision for, and review of, electrical and magnetic geophysical methods, *Hydrol. Processes*, *22*(18), 3604-3635.

Rodgers, P., C. Soulsby, and D. Tetzlaff (2005), Using stable isotope tracers to assess hydrological flow paths, residence times and landscape influences in a nested mesoscale catchment, *Hydrol. Earth Syst. Sci.*, *9*, 139-155.

Roering, J. J., J. W. Kirchner, and W. E. Dietrich (2005), Characterizing structural and lithologic controls on deep-seated landsliding: Implications for topographic relief and landscape evolution in the Oregon Coast Range, USA, *Geological Society of America Bulletin*, *117*(5-6), 654-668.

Sayama, T., and J. J. McDonnell (2009), A new time-space accounting scheme to predict stream water residence time and hydrograph source components at the watershed scale, *Water Resour. Res.*, *45*.

Sayama, T., J. J. McDonnell, A. Dhakal, and K. Sullivan (2011), How much water can a watershed store?, *Hydrol. Processes*, n/a-n/a.

Schlosser, P., M. Stute, H. Dorr, C. Sonntag, and K. O. Munnich (1988), TRITIUM HE-3 DATING OF SHALLOW GROUNDWATER, *Earth and Planetary Science Letters*, *89*(3-4), 353-362.

Sidle, R. C., S. Noguchi, Y. Tsuboyama, and K. Laursen (2001), A conceptual model of preferential flow systems in forested hillslopes: evidence of self-organization, *Hydrol. Processes*, *15*, 1675-1692.

Snavely, P. D., H. C. Wagner, and N. S. MacLeod (1964), Rythmic-bedded eugeosynclinal deposits of the Tyee Formation, Oregon Coast Range, in *Symposium on cyclic sedimentation: Kansas Geological Survey, Bulletin 169, pp. 461-480*, edited by D. F. Merriam.

Solomon, D. K., and P. G. Cook (2000), $^3H$ and $^3He$, in *Environmental Tracers in Subsurface Hydrology*, edited by P. Cook and A. L. Herczeg, pp. 397-424, Kluwer Academic Publishers, Boston.

Solomon, D. K., D. P. Genereux, L. N. Plummer, and E. Busenberg (2010), Testing mixing models of old and young groundwater in a tropical lowland rain forest with environmental tracers, *Water Resour. Res.*, *46*.

Soulsby, C., D. Tetzlaff, and M. Hrachowitz (2009), Tracers and transit times: windows for viewing catchment scale storage?, *Hydrol. Processes*, *23*(24), 3503-3507.

Soulsby, C., D. Tetzlaff, N. van den Bedem, I. A. Malcolm, P. J. Bacon, and A. F. Youngson (2007), Inferring groundwater influences on surface water in montane catchments from hydrochemical surveys of springs and streamwaters, *J. Hydrol.*, *333*(2-4), 199-213.

Spence, C. (2010), A Paradigm Shift in Hydrology: Storage Thresholds Across Scales Influence Catchment Runoff Generation, *Geography Compass*, *4*(7), 819-833.

Stewart, M. K., U. Morgenstern, and J. J. McDonnell (2010), Truncation of stream residence time: how the use of stable isotopes has skewed our concept of streamwater age and origin, *Hydrol. Processes*, *24*(12), 1646-1659.

Swanson, F. J., and M. E. James (1975), Geology and geomorphology of the H.J. Andrews Experimental Forest, western Cascades, Oregon.*Rep. Res. Pap. PNW-188*, 14 pp, U.S. Department of Agriculture, Forest Service, Pacific Northwest Forest and Range Experiment Station, Portland, OR.

Tetzlaff, D., J. Seibert, and C. Soulsby (2009), Inter-catchment comparison to assess the influence of topography and soils on catchment transit times in a geomorphic province; the Cairngorm mountains, Scotland, *Hydrol. Processes*, *23*(13), 1874-1886.

Thorson, T. D., S.A. Bryce, D.A. Lammers, A.J. Woods, J.M. Omernik, J. Kagan, D.E. Pater, and J.A. Comstock (2003), Ecoregions of Oregon, *U.S. Geological Survey, Reston, VA.*

Torres, R., W. E. Dietrich, D. R. Montgomery, S. P. Anderson, and K. Loague (1998), Unsaturated zone processes and the hydrologic response of a steep, unchanneled catchment, *Water Resour. Res.*, *34*(8), 1865-1879.

Uchida, T., J. J. McDonnell, and Y. Asano (2006), Functional intercomparison of hillslopes and small catchments by examining water source, flowpath and mean residence time, *J. Hydrol.*, *327*, 627-642.

Wagener, T., M. Sivapalan, P. Troch, and R. Woods (2007), Catchment classification and hydrologic similarity, *Geography Compass*, *1*.

Western, A. W., R. B. Grayson, and G. Bloschl (2002), Scaling of soil moisture: A hydrologic perspective, *Annual Review of Earth and Planetary Sciences*, *30*, 149-180.

Weyman, D. R. (1973), Measurements of the downslope flow of water in a soil, *J. Hydrol.*, *20*, 267-288.

Whipkey, R. Z. (1965), Subsurface storm flow from forested slopes, *Bull. Int. Assoc. Sci. Hydrol.*, *2*, 74-85.

Wilson, C. J., and W. E. Dietrich (1987), The contribution of bedrock groundwater flow to storm runoff and high pore pressure development in hollows, in *Erosion and Sedimentation in the Pacific Rim*, edited, IAHS Publication No. 165, Corvallis, OR.

Wondzell, S. M. (2006), Effect of morphology and discharge on hyporheic exchange flows in two small streams in the Cascade Mountains of Oregon, USA, *Hydrol. Processes*, *20*(2), 267-287.

Table 3.1. Mean deuterium composition and standard deviation of precipitation, streamwater, and groundwater.

| Location | Mean $\delta^2H$ [‰] | $\delta^2H$ Standard Deviation | n |
|---|---|---|---|
| Amount-weighted precipitation | -50.6 | 15.1 | 170 |
| NB-12 | -49.6 | 1.8 | 108 |
| NB-86 | -50.1 | 2.2 | 179 |
| Well-1DP | -51.2 | 0.4 | 8 |
| Well-2DP | -52.0 | 1.4 | 2 |
| Well-3HS | -49.7 | Na | 1 |
| Well-8HS | -50.6 | 1.5 | 2 |
| Well-9HS | -48.3 | 0.8 | 2 |
| Well-10FP | -50.5 | 1.2 | 3 |
| Well-11FP | -50.8 | 0.4 | 4 |
| Well-16SF | -48.4 | 1.9 | 12 |

Table 3.2. Soil water mean transit time modeling results. Maximum likelihood estimates (mle) for the dispersion parameter (*$D_p$*) and mean transit time ($\tau_m$) parameter of the dispersion flow model, parameter uncertainties, and Nash-Sutcliffe efficiencies (NSE) are presented. Both locations represent 0.5 m soil depths.

| Site | $D_{p\text{-mle}}$ | $D_{p\text{-}10/90\ \%}$ | $\tau_{m\text{-mle}}$ [d] | $\tau_{m\text{-}10/90\ \%}$ [d] | $NSE_{mle}$ |
|---|---|---|---|---|---|
| Midslope | 0.06 | (0.05/0.48) | 132 | (96/134) | 0.57 |
| Toeslope | 0.59 | (0.54/0.66) | 63 | (59/69) | 0.79 |

Table 3.3. Tritium-based mean transit time results with deuterium-based transit times provided for comparison.

| Location | Model | Exponential Fraction, f | $^3H$-based MTT [yrs] | $\delta^2H$-based MTT [yrs] |
|---|---|---|---|---|
| NB-12 | EPM | 0.7 | 6.8 (5.1/8.5) | 5.0 (4.0/8.7) |
| NB-86 | EPM | 0.7 | 7.7 (6.0/9.4) | 4.0 (3.5/4.9) |

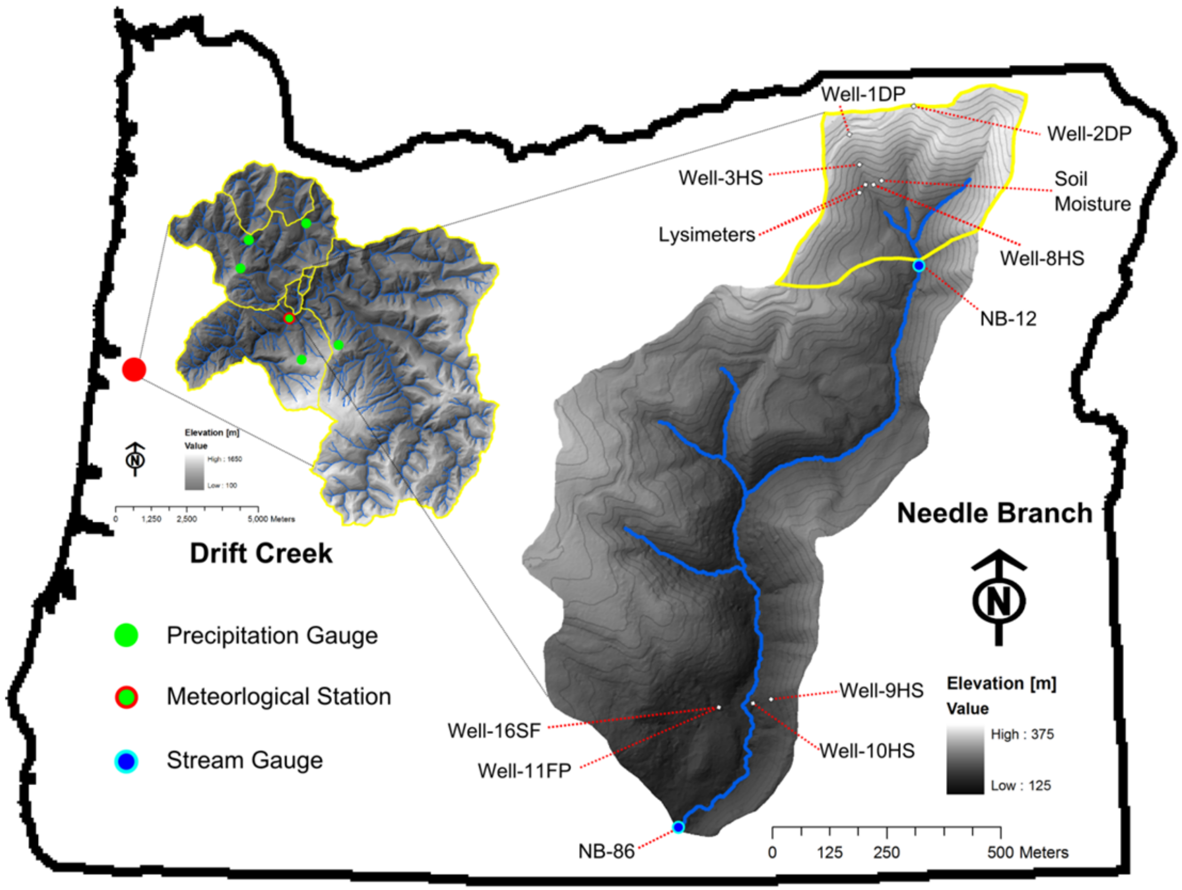


Figure 3.1. Study site map. Contour intervals are 10 m.

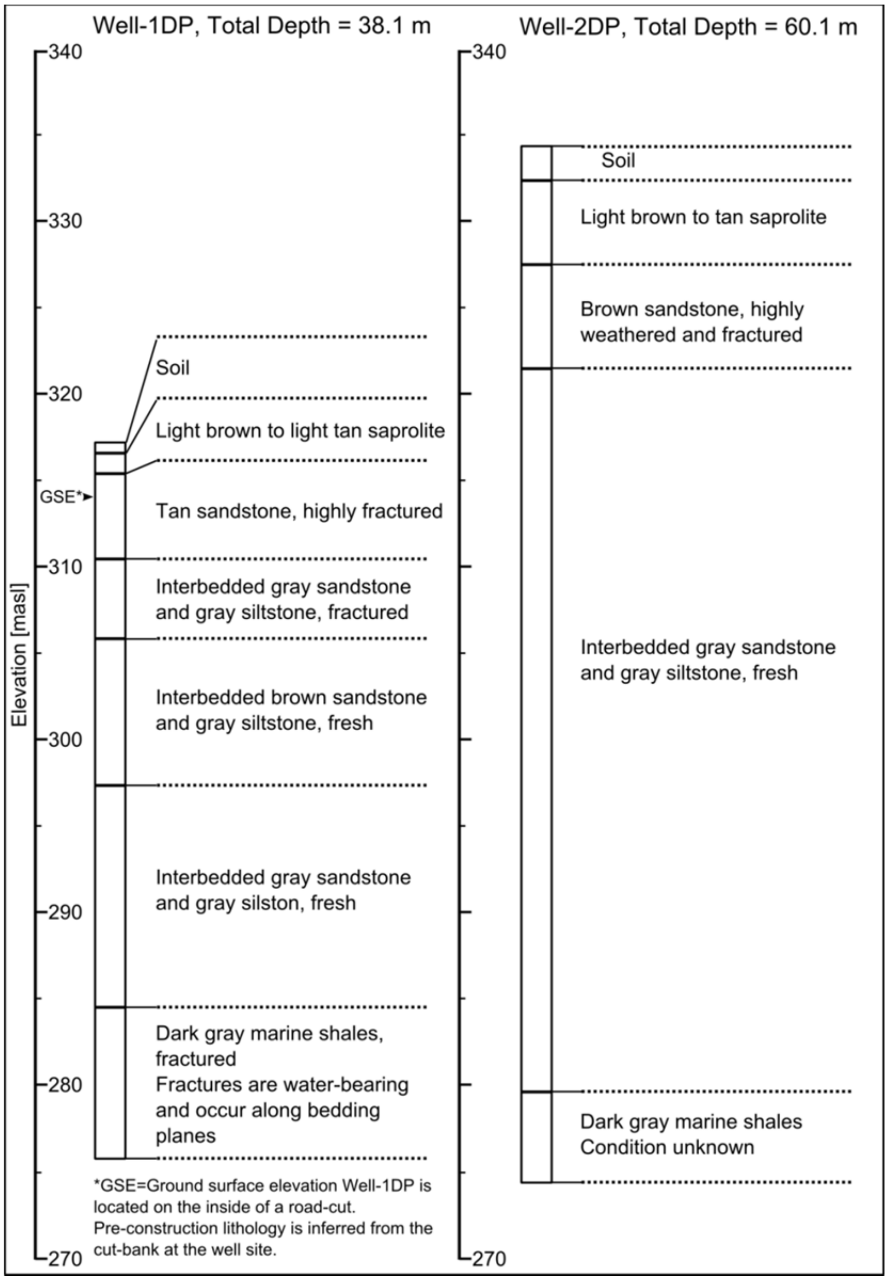


Figure 3.2. Boring logs from Well-1DP and Well-2DP.

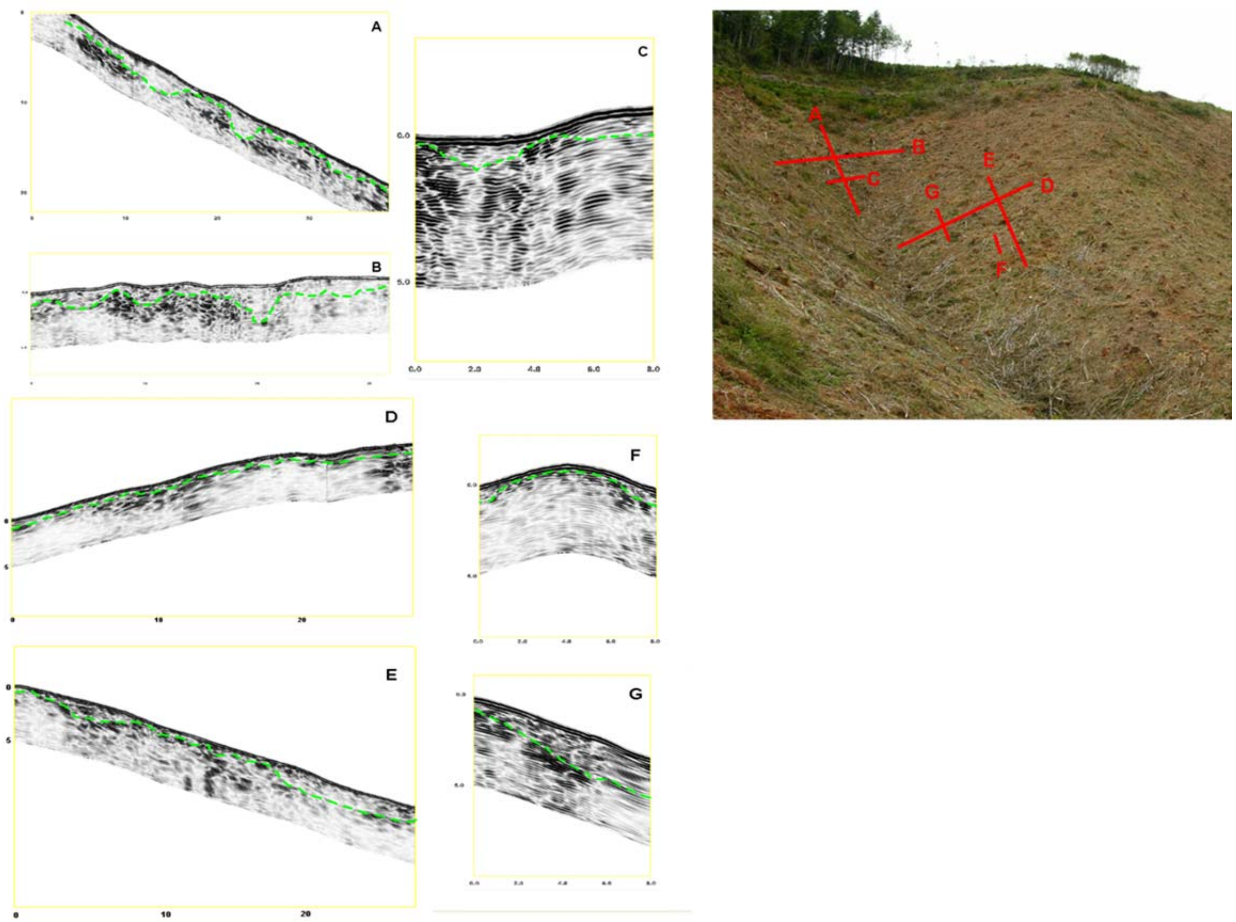


Figure 3,3. Ground penetrating radar profiles for transects in the western-most sub-catchment of NB-12, A-F, as depicted in the inset photo. Approximate soil-bedrock interface shown as a dashed green line.

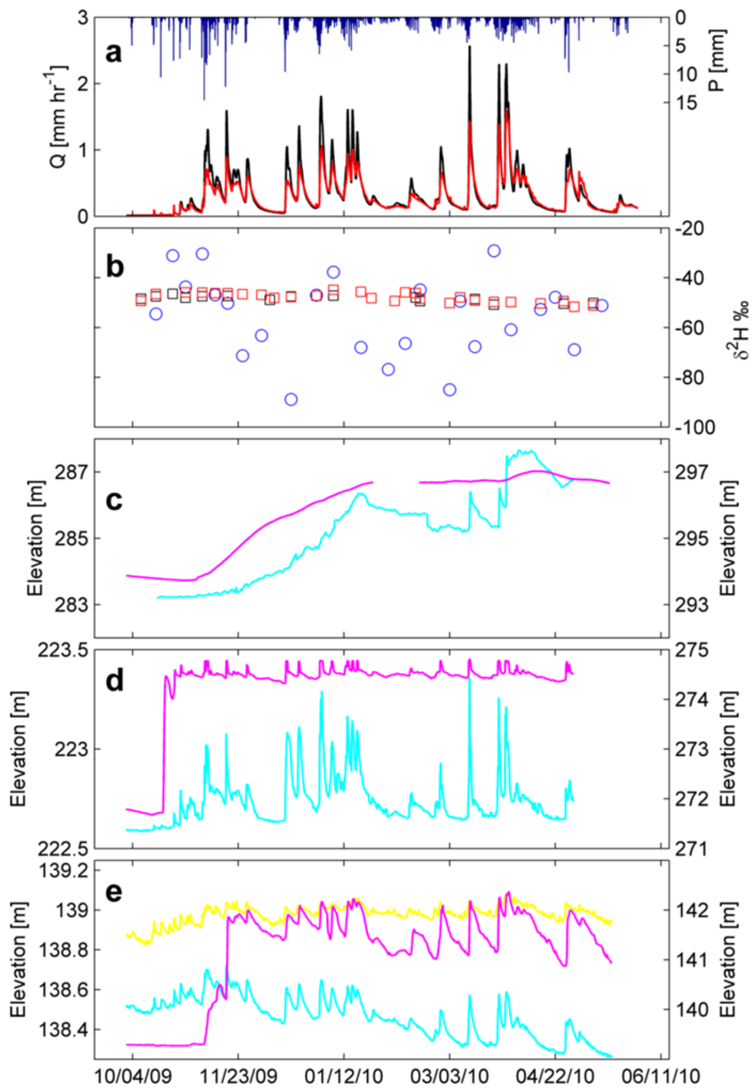


Figure 3.4.Hydrometrics and tracer dynamics measured from 01 October 2009 through 30 May 2010. Precipitation (blue) and specific discharge for NB-12 (black) and NB-86 (red) is shown in (a). Deuterium composition of precipitation (blue circles), NB-12 streamwater (black squares), and NB-86 streamwater (red squares) is shown in (b). Water level elevations Well-1DP (cyan, left axis) and Well-2DP (pink, right axis) are shown in (c). Water level elevations for Well-3HS (pink, right axis) and Well-8HS (cyan, left axis) are shown in (d). Water level elevations for Well-9HS (pink, right axis), Well-10FP (cyan, left axis), and Well-11FP (yellow, left axis) are shown in (e).

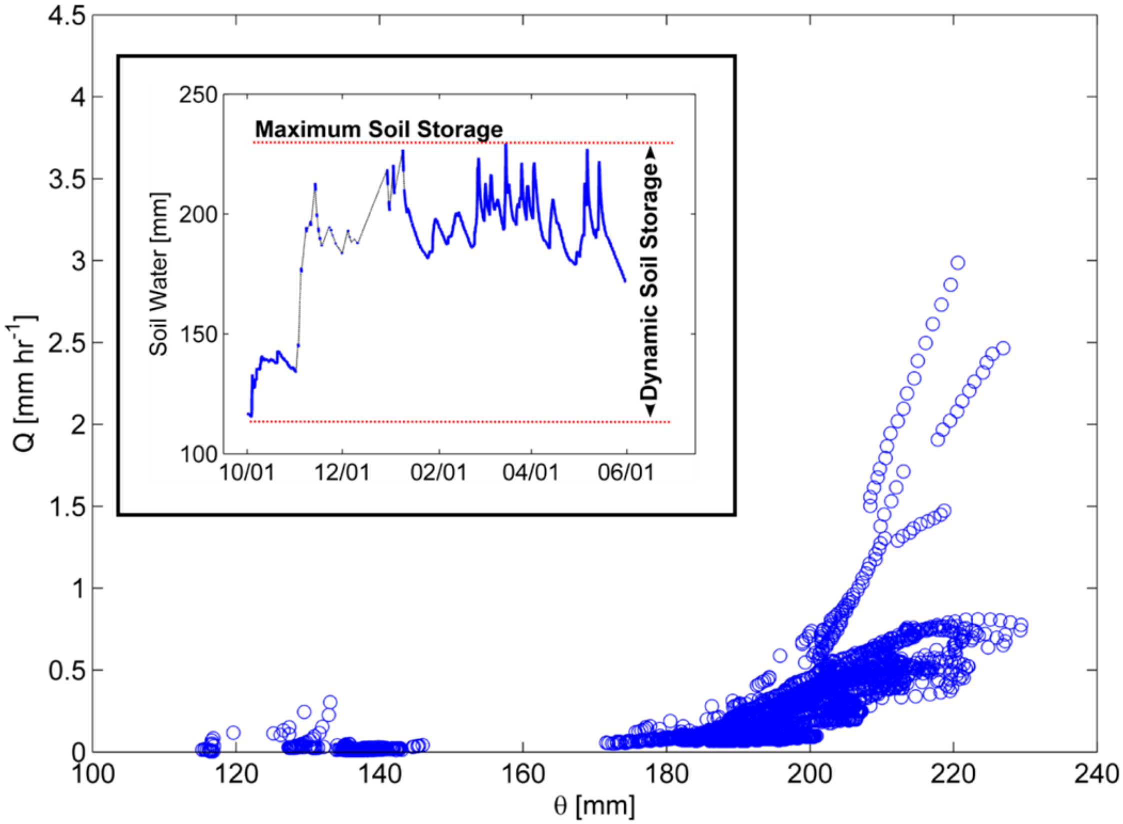


Figure 3.5. Stream discharge as a function of soil water content. Water depth is estimated using the integration of measurements obtained from four soil moisture sensors distributed within a 1 m deep soil profile (0.15, 0.35, 0.65, and 0.85 m depths). The inset shows the times series of soil water depth calculated during the course of the wet-up and wet periods at a representative hillslope profile that was used to construct the main figure. The dashed black line in the inset represents values interpolated to fill data gaps (not used in the analysis shown in the main figure).

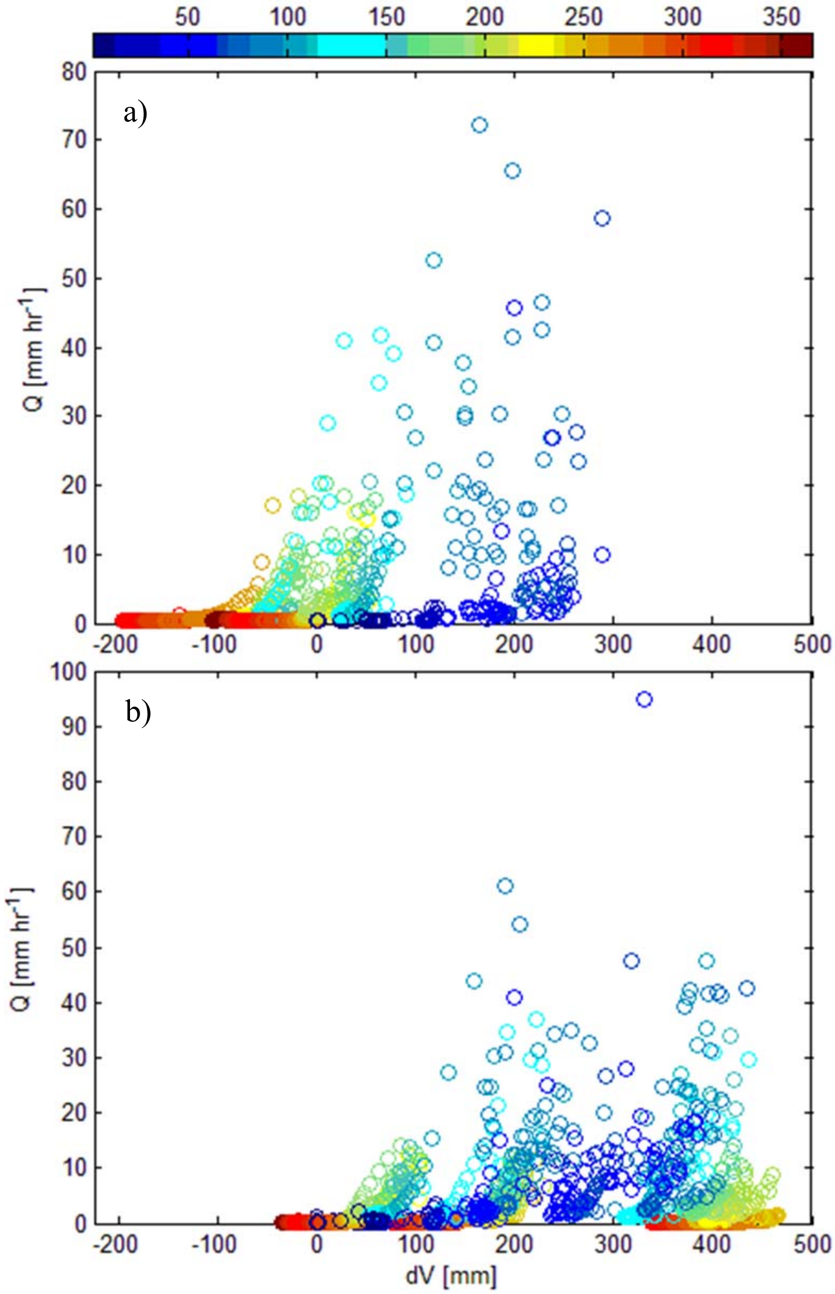


Figure 3.6. Discharge as a function of change in storage (dV) since the beginning of the water year for NB-12 (a) and NB-86 (b). Marker color represents the day of the water year.

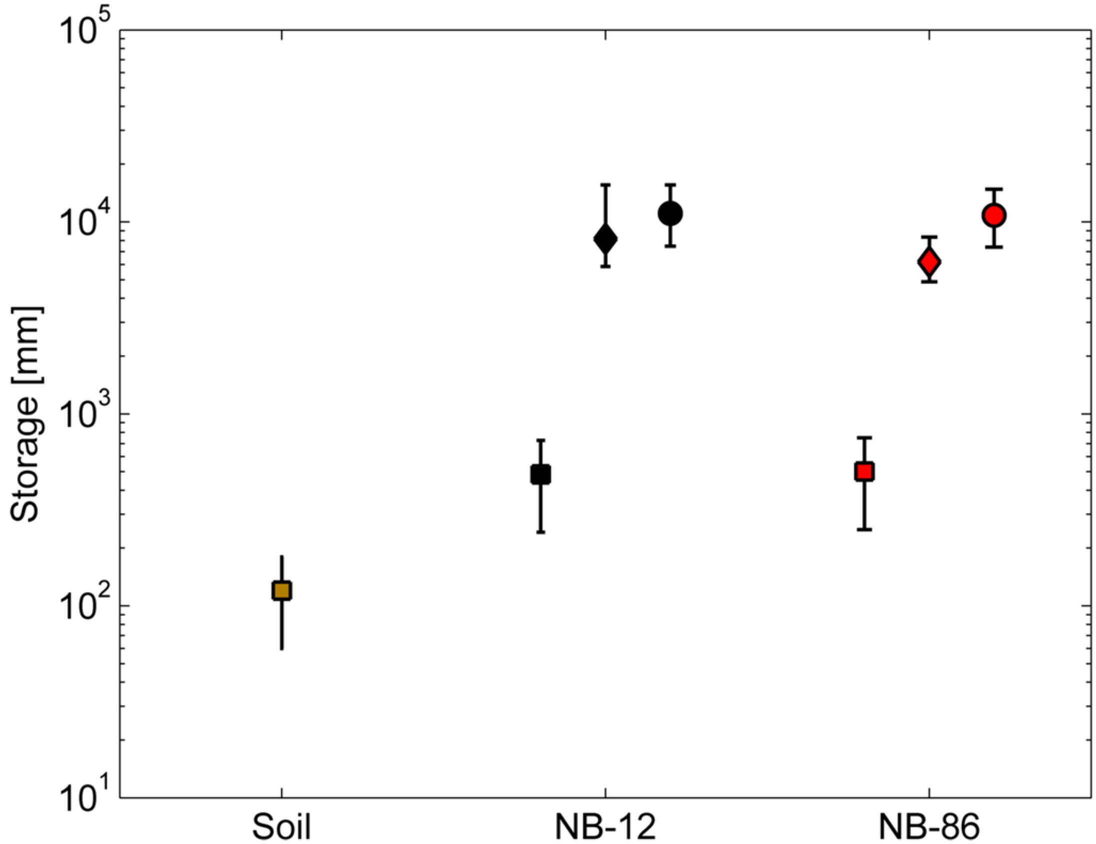


Figure 3.7. Comparison of dynamic (squares) and total storage (deuterium-based estimates are represented with diamonds and tritium-based estimates are represented with circles) for the soil profile (brown), NB-12 (black), and NB-86 (red).

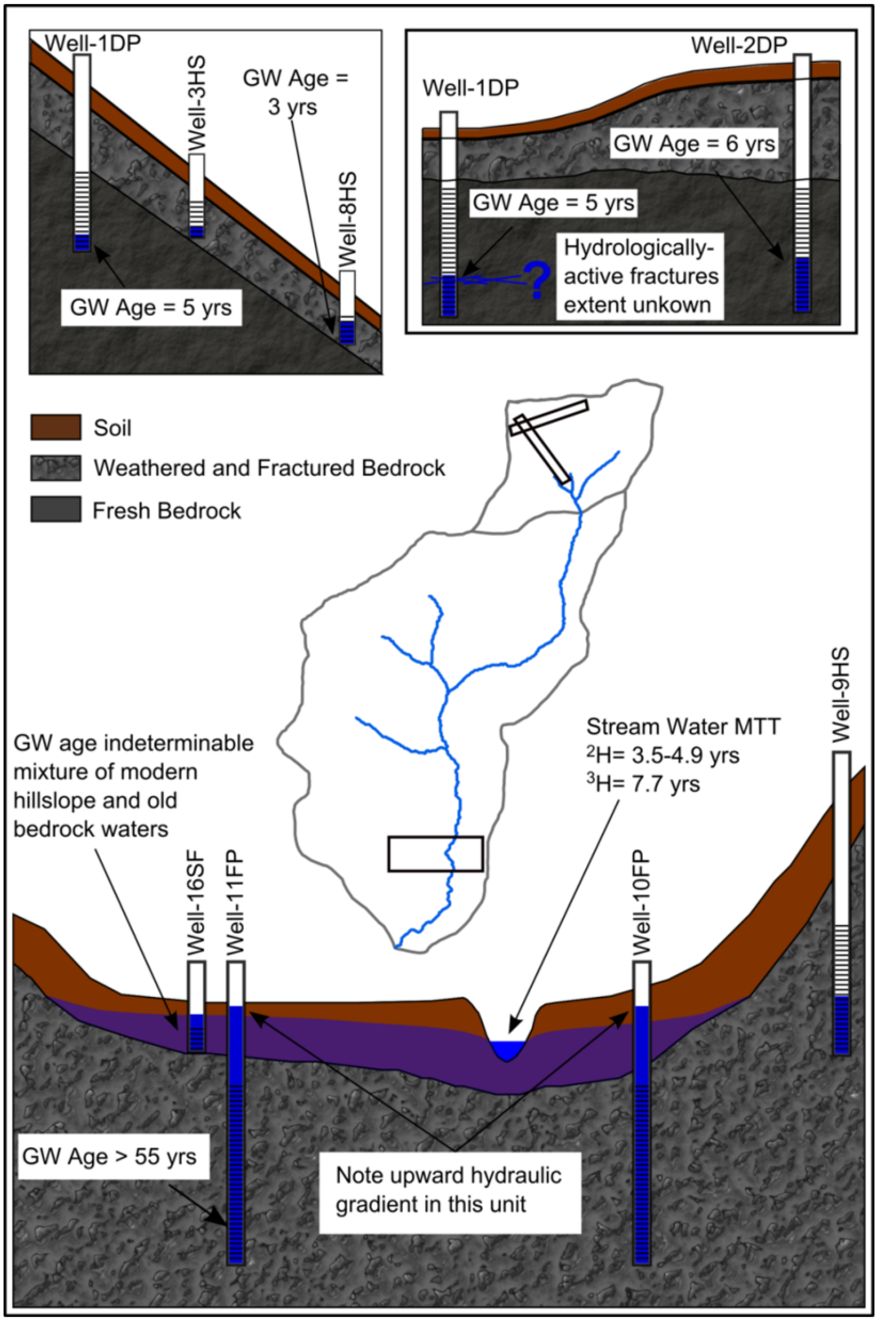


Figure3.8. Diagram showing groundwater ages.

# Where Are Diel Streamflow Fluctuations Generated?: A Test of the Near-Stream Zone of Influence Hypothesis

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**INTRODUCTION**

Forest vegetation interacts with the hydrologic cycle via two primary pathways: canopy interception and plant transpiration [*Hewlett*, 1982]. While canopy interception of water (and subsequent throughfall or evaporation) is purely abiotic, the loss of water through transpiration is a physiologically-mediated process. Since transpired water is typically extracted from the subsurface by the plant's root system, forest vegetation plays an active role in regulating catchment storage during the growing season in temperate zones. Despite a well-developed understanding of water uptake by plant roots at the cellular level [*Raats*, 2007; *Steudle*, 2000], a large degree of ambiguity surrounds the conceptualization of how forest vegetation interacts with subsurface storage components at the catchment scale [*Asbjornsen et al.*, 2011]. Hence, an enduring question for forest hydrologists, is "how, where, and under what conditions do forests exert influence on the low-flow hydrology of forested catchments?"

Physiologically-mediated, transpiration is externally driven by climate variables affecting photosynthetic rates (solar radiation) and evaporative demand (vapor pressure deficit, referred to hereafter as VPD). Therefore, when considered on the sub-daily timescale, transpiration-induced influence on catchment storage is periodic with peaks occurring near midday (lagged behind peak solar irradiance and VPD) with rates approaching zero during the night. This periodic influence has long been linked to diurnal suppression of groundwater levels and stream discharge in instrumented catchments around the world [*Meinzer*, 1927; *Troxell*, 1936; *White*, 1932; *Wicht*, 1941]. However, not all diel fluctuations are driven by transpiration [*Lundquist and Cayan*, 2002]; diel cycles in the magnitude of radiation inputs can create fluctuations as a result of daily thawing and freezing of the snowpack in snow-dominated systems or in losing streams where streambed infiltration rates are controlled by viscosity and hydraulic conductivity, both temperature-dependent [*Constantz*, 1998]. Where fluctuations are driven by transpiration, interactions between the forest and different components/compartments of the hydrologic system are embedded in the fluctuating diel signal. The advent of relatively low-cost instrumentation capable of high-precision, near-continuous water level measurements has renewed interest in exploring the information content of diel fluctuations [*Gribovszki et al.*, 2010].

While most diel fluctuation studies to date have focused on using characteristics of the measured signal to estimate riparian evapotranspiration rates [*White*, 1932], relatively few have been able to identify the mechanisms and processes controlling the signal. Indeed, there have been fundamental disagreements in the literature about such mechanisms [*Wondzell et al.*, 2007; 2010]. The Pacific Northwest of the USA has been a particularly useful place for such process-based research because the Mediterranean climate and associated seasonal drought that is coincident with the growing season creates a system where biology and hydrology are out of phase [*Brooks et al.*, 2010]. This allows for evaluation of the diel streamflow signal with little to no interference by precipitation events.

Several such studies have originated from the HJ Andrews Experimental Forest (HJA) located in the central Oregon Cascades Range. The first intensive examination of forest-stream coupling at the HJA was a study by *Bond et al.* [2002] that proposed that the diel signal was created only by trees located close to the stream. *Bond et al.* [2002] used onsite estimations of transpiration, scaled from sapflow measurements, to show that only a small area of transpiring vegetation was necessary to account for the daily suppression of stream discharge at the 100 ha HJA-WS01 catchment. They also found that the amplitude of the diel streamflow fluctuations decreased with time and became increasingly lagged relative to the transpiration signal (what was assumed to be driving the streamflow fluctuations). They hypothesized that this was due to a shrinking “zone of influence” that corresponded to a decreasing water table elevation in the riparian zone. Thus, as the dry period progressed, fewer and fewer trees were “connected” to the shallow, alluvial aquifers feeding stream baseflow. *Wondzell et al.* [2007] provided an alternative hypothesis for explaining the decreasing amplitude and increasing time lag relative to transpirational demand observed at HJA-WS01. They hypothesized that because decreases in discharge volumes correspond to decreases in water velocities, velocity-dependent signal attenuation within the channel explain the observed decreases in amplitude and increases in time lag relative to transpiration. However, in a follow-up paper, *Wondzell et al.* [2010] found that their initial conceptual model, concerning only the riparian zone and stream channel network, was too simplistic and that both riparian and hillslope inputs must be necessary to fully explain the generation of diel streamflow fluctuations. *Barnard et al.* [2010] showed in the nearby HJA-WS10 catchment that indeed diel fluctuations were observed in hillslope seepage. They found that hillslope seepage fluctuations were dependent on soil moisture content, but their methodology did not allow spatial discretization of where the diel seepage signal was generated on the hillslope. This catchment-scale question of "where" forest vegetation exerts influence on streamflow is the fundamental next step question for the HJA, the PNW and forested catchments in general.

Here, we attempt to identify where diel fluctuations are generated using a field-based rejectionist approach. We test directly the *Bond et al.* [2002] "near-stream" hypothesis by experimentally removing trees within the near-stream zone in two 2.5 ha headwater catchments in Western Oregon, 140 km west of HJA-WS01 (the site of the original *Bond et al.* [2002] study). To our knowledge, this is the first time such a manipulative experiment has been carried out for this purpose (although *Dunford and Fletcher* [1947] and *Rowe* [1963] performed riparian vegetation manipulation to explore water yield augmentation questions). In addition to the manipulative experiment, we use the stable isotopes of hydrogen and oxygen ($\delta^2$H and $\delta^{18}$O, respectively) as tracers to provide further rejection criteria for testing the "near-stream" hypothesis. Stable isotopes have become invaluable tools for elucidating ecohydrologic processes [*Asbjornsen et al.*, 2011]. At natural abundance levels, the isotopic composition of $\delta^2H$ and $\delta^{18}O$ in water can be physically influenced in a deterministic way by equilibrium and kinetic fractionation processes [*Gat and Tzur*, 1967]. As a result, water from different precipitation events, or storage components, often have unique isotopic signatures. Since water uptake by tree roots is generally an advective process, no fractionation occurs as water passes from the soil into the roots [*Ehleringer and Dawson*, 1992]. Therefore, if different subsurface water stores accessible to trees are isotopically unique (i.e. soil water, groundwater, and stream water), xylem water sources can be evaluated in a simple, dual isotope mixing-analysis framework. We utilize this methodology to test if the trees expected to be influencing streamflow are in fact utilizing water sources contributing to streamflow for transpiration.

In combination, our manipulative experiment and isotope tracing tests four elements of the *Bond et al.* [2002] "near-stream" hypothesis:

1) Do near-stream trees control the diel fluctuations observed in growing-season stream discharge records?

2) If diel fluctuations are not generated solely by trees in the near-stream zone, is the signal generated within the boundaries of the catchment?

3) If a tree is active in generating diel fluctuations, is it also accessing a water source that is contributing to streamflow?

4) Do trees switch from using water that is connected to streamflow to using water not connected to streamflow as antecedent moisture declines?

## STUDY SITE

Our research catchments are located in the central Oregon Coast Range (44.52° N 123.85° W) (Figure 4.1), 140 km west of HJA-WS01. This area lies within the Mid-

Coastal Sedimentary Ecoregion [*Thorson*, 2003] which is a highly-dissected mountainous area, characterized by short, steep slopes which give rise to medium- to high-gradient streams. Upland soils are loams to gravelly loams (mesic Alic Hapludands and mesic Andic Humudepts) that average 1m depth and are classified as well- to very well-drained [*Corliss*, 1973]. Valley bottom soils are silt loams (isomesic Fluvaquentic Humaquepts), which average 2 m depth and are classified as somewhat poorly drained. The underlying bedrock is the Eocene-aged Tyee Formation of rhythmic-bedded layers of marine-derived sandstones and siltstones [Snavely et al., 1964]. Mean annual precipitation at our site is approximately 2,440 mm. On average, greater than 85% of the annual precipitation occurs from October through April in "long-duration, low-to-moderate intensity frontal storms" [*Harr*, 1976]. Snow, while occurring occasionally, does not usually accumulate; studies in the past have neglected snow as part of the precipitation record [*Harris*, 1977].

We conducted our experiments and sampling in an 86 ha catchment (NB-86; see Figure 1) during the 2008 and 2009 growing seasons (approximately May through September). At the beginning of the study period, canopy vegetation was dominated by 41-yr-old Douglas-fir (*Pseudotsuga menziesii*) on the hillslopes with red alder (*Alnus rubra*) of the same age-class inhabiting both riparian and some hillslope areas. We used a nearby (<2 km to the north), 202 ha catchment (Flynn Creek) with a naturally regenerating mixed conifer-hardwood forest, also predominantly red alder and Douglas-fir, as an external, but nearby control.

## METHODS

### *Near-stream zone of influence experiment*

We designed an experiment to test the near-stream "zone of influence" hypothesis put forward by *Bond et al.* [2002]. The goal of the experiment was to shut down the diel fluctuation in stream discharge by felling trees along the stream corridor. We took advantage of a planned industrial forest harvest operation at our research catchments to employ this manipulative experiment of near-stream tree felling prior to the remaining trees in the harvest unit being felled.

The Oregon Forest Practices Act requires vegetated buffers to be retained adjacent to fish-bearing stream reaches (Oregon Administrative Rules 629-635-0000), so our experiment was limited to the non-fish-bearing, $1^{st}$-order streams in the headwaters of NB-86. Of the three $1^{st}$-order catchments, we selected the two that flowed through the previous growing seasons as for our experiment (Figure 4.1); Catchment A (2.4 ha) and Catchment B (2.6 ha). The streams draining these two catchments were steep and incising a v-shaped valley, resulting in fairly simple catchment geomorphology consisting of planar hillslopes and a hillslope hollow. This differs from the HJA-WS01 stream where *Bond et al.* [2002] worked that drained a 96 ha catchment and, therefore, had additional geomorphic complexity as a function of its larger size. The additional complexity at HJA-WS01 included an alluvial plain with an underlying alluvial aquifer [*Wondzell*, 2006] which provided a direct, intuitive connection to the surface water system. Nonetheless, pilot data from our stream gauges showed that diel discharge fluctuations were present even in our extreme headwater streams where opposing hillslopes converge directly into the stream channel (Figure 4.2).

Stream discharge was measured at Catchment A and Catchment B using calibrated 15° v-notch weirs retrofitted to trapezoidal flumes (Model Small 60° V, Plasti-fab, Inc., Tualatin, OR) where stream stage was measured with capacitance rods (Model WT-HR, Tru-track, Inc.) on 10-min intervals (precision= 1.0 mm). A broach-crested, compound v-notch weir, installed by the U.S. Geological Survey, was used to measure stream discharge at our 202 ha control catchment, Flynn Creek. Stream stage was measured on 10-min intervals with a vented pressure transducer (Model PDCR 1830, GE Druck, Inc., precision=0.1 mm).

We felled the trees in the riparian corridor of Catchment A first (referred to hereafter as Experiment A), using the immediately adjacent and approximately equal-area Catchment B as the control. We replicated the experiment in Catchment B (Experiment B), using the nearby Flynn Creek as the control catchment. In each case, the felled trees were left in place until the remaining trees in both experimental catchments were felled and subsequently removed from the site (as part of the industrial forest harvest operation) following the post-treatment data collection period for Experiment B.

The number of trees felled in each experiment was determined by estimating the area of transpiring vegetation necessary to account for the volume of water "missing" from the discharge record, following the method of *Bond et al.* [2002]. "Missing" discharge was calculated by fitting a line from the peak hourly discharge on $day_i$ to peak hourly discharge on $day_{i+1}$ and then summing the difference between the fitted line and the measured discharge over the period for which they diverged (e.g. *Bond et al.* [2002] and *Barnard et al.* [2010]). To determine the area of transpiring vegetation necessary to account for the "missing" water, we used a transpiration rate of 2 mm $day^{-1}$. This value is within the range of early growing season values (June) reported by *Moore et al.* [2004] from HJA-WS01, where the forest age and species distribution is nearly identical to our experimental catchment (mixed 40-yr-old red alder and Douglas-fir in the riparian zone at the time of the study). Although they estimated daily transpiration values greater than 3 mm in June 2000 (as much as 3.6 mm $day^{-1}$), we chose to use 2 mm $day^{-1}$ as a conservative estimate for our calculations so that we would remove at least as much area of transpiring vegetation necessary to account for the "missing" water in the daily discharge fluctuation.

We measured sap velocities to use as a direct, but relative measure of transpiration magnitude and timing. In 2008, sap velocities were measured in 16 dominant and co-dominant red alder and Douglas-fir trees (8 for each species) using constant-heat sapflow sensors [*Granier*, 1987]. The outer 20 mm of sapwood (0-20 mm radial depth) was measured with a single 20 mm sensor installed in each tree. For the experiments, the sensor network was reduced to 4 representative trees (2 red alder and 2 Douglas-fir) located just outside of the expected tree removal area in Catchment A (but not on the boundary so as to minimize any edge effects). Following the felling experiment, the sensor network was relocated to the lower catchment, outside of the harvesting operation (Figure 4.1), where sensors were installed in 3 trees in each of four plots; located on the hillslope, toeslope, stream bank, and alluvial plain (Figure 4.3; 12 sensors total). For all cases, measurements were taken every 10 s and averaged and stored to a datalogger (CR10X, Campbell Scientific, Inc.) every 10 min. Vapor pressure deficit (VPD; kPa), determined from meteorological measurements made in a clearing at the base of the Needle Branch catchment, was used as a surrogate for sap velocity when those measurements were unavailable.

*Statistical approach for detecting change*

We assessed the effectiveness of our experimental tree felling in shutting down the diel streamflow fluctuations using the magnitude and timing of the fluctuation as metrics for detecting change. We used the daily amplitude of the diel fluctuation (calculated as the difference between the maximum and minimum discharge measured each day) to characterize the fluctuation magnitude. We evaluated the change in the timing of the signal relative to the peak sap velocity; using cross-correlation analysis to identify the time lag at the point of the most negative correlation for each diel cycle (essentially determining the lag which best aligned peak sap velocity with the minimum stream discharge).To detect change, we compared the treatment-control ratio of each metric (fluctuation magnitude and fluctuation timing) for the pre-treatment and post-treatment periods (following Jones and Post [2004]). We used two-sample t-tests to test the null hypothesis that the mean treatment-control relationship for the pre-treatment was equal to the mean treatment-control relationship of the post-treatment period. We used a significance level of 0.05 as our requirement for rejecting the null hypothesis ($\alpha$=0.05).

*Testing for influence beyond the near-stream zone*

We used stream discharge and groundwater levels to assess the persistence of diel fluctuations following the felling and removal of all trees onsite. Stream discharge was measurements are described in the previous section. Diel fluctuations in groundwater following the harvest operation was monitored in two hillslope wells installed into shallow bedrock in Catchment A using the portable drilling system described by Gabrielli and McDonnell (2011). Well-7HS (HS=hillslope) was installed to a total depth of 6.74 m and Well-8HS was 5.67 m total depth (Figure 4.1). The wells are all constructed of 25.4 mm inside-diameter (ID) polyvinylchloride (PVC). A 101.6 mm ID PVC outer casing was installed to the depth of the saprolite-bedrock transition. The inner casings were custom-made for each well so that the screened interval was sealed to a depth of 0.5 to 1 m below the saprolite-bedrock transition (3.05 and 2.47 m for Well-7HS and Well-8HS, respectively). The screen section was created using approximately 0.5 mm horizontal slots; the slotted section was covered with two layers of fiberglass mesh screen to prevent fine particles from entering the well. The annulus between the inner casing and the borehole wall was not large enough to allow backfilling with sand to an accurate, evenly packed depth, so the seal was created by positioning a rubber gasket on the outside of the inner casing such that it fit snuggly against the walls of the borehole at the desired sealing depth. Bentonite was backfilled on top of the gasket to the top of the outer casing to finalize the seal (approximately 0.1 m above the ground surface). Water levels were measured on 10-min intervals with water level loggers (Model U20, Onset Computer Corporation, Inc., Pocasset, MA) and the records were corrected for barometric effects using an onsite barometric pressure measurement made with the same sensor.

*Isotope tracing to determine tree water source*

We used the stable isotopes of water, $\delta^2H$ and $\delta^{18}O$, to test whether the water sources trees utilized for transpiration were also contributing to streamflow and to determine if trees using water sources contributing to streamflow under wet antecedent wetness conditions switched to other water sources when antecedent conditions were drier. Four sample plots were established along a hillslope catena so that water sources could be discriminated based on landscape position—hillslope, toeslope, streambank, and alluvial plain (Figure 4.3). Groundwater wells were installed at each plot to provide access for sample collection and to monitor water table elevations. Antecedent wetness was estimated using an antecedent precipitation index (API) calculated using precipitation records measured at a tipping bucket rain gauge (Model TR-525M, Texas Electronics, Inc., Dallas, TX) located in a clearing near the NB-86 stream gauge (Figure 4.1). The index was calculated as,

$$API_M = API_M(t-1)e^{\left(\frac{\ln(0.5)}{M}\right)} + P_i(t)e^{\left(\frac{\ln(0.5)}{2M}\right)} \tag{1}$$

where $t$ is time, $Pi(t)$ is the precipitation that occurs from time $t-1$ to $t$ ,and $M$ is the half-life period indicating that precipitation decays to one-half its original value $M$ time-steps after its occurrence [*Descroix et al.*, 2002]. We calculated $API_7$, $API_{14}$, and $API_{30}$ to provide a sense of wetness based on different window sizes (i.e. $M$ values).

Well-9HS (total depth=7.45 m, seal depth=4.27 m) was located at the hillslope plot and installed, constructed, and monitored following the same procedures outlined for Well-7HS and Well-8HS. Well-11SF (SF=surficial, total depth=2.40 m), Well-14SF (total depth=1.25 m), and Well-16SF (total depth=1.12 m) were installed to the soil-bedrock interface at the toeslope, streambank, and alluvial plain plots, respectively. The SF wells were constructed of 25.4 mm (ID) PVC, screened over the lower 0.40 m section (using the same construction as the HS wells), and installed in hand-augered holes. The annulus was backfilled with packed sand to 0.25 m above the screened section and sealed with bentonite to the ground's surface. Groundwater levels were monitored in the SF wells using capacitance rods (previously described).

Xylem water, soil water, stream water, and groundwater were collected on a monthly basis throughout the 2009 growing season (May through September). Stream water and groundwater were sampled directly (all wells were purged prior to sampling). Soil water and xylem water were extracted from soil and xylem cores using cryogenic vacuum distillation [*Ehleringer and Osmond*, 1989]. Soil was sampled at 10, 30, 50, 90, 120, and, where possible, 150 cm depths from a single auger-hole in each plot for all sampling events (depths are from the mineral soil surface). For the September 2009 event, soil was sampled from the surface of the mineral soil in addition to the regular depths. Xylem cores were extracted from three trees in each plot using a 12mm increment borer (Haglöf Sweden). All samples were sealed immediately in glass vials, taking care to minimize contact time with the ambient air to prevent fractionation.

Water samples were analyzed using off-axis integrated cavity output laser spectroscopy on a Los Gatos Research Liquid Water Isotope Analyzer (LWIA-24d, Los Gatos Research, Inc.) at the Oregon State University Water Isotope Collaboratory (OSU-WIC). Recent work by *West et al.* [2010] shows that dissolved organic compounds in extracted xylem water can interfere with the absorbance spectra utilized by laser spectroscopy methods and thus affect the accuracy of the measurement; based on this development, we reanalyzed our xylem water samples using isotope ratio mass spectrometry (IRMS) on a Delta Plus XL mass spectrometer (ThermoFinnigan; analyses conducted by the Center for Stable Isotope Biogeochemistry at the University of California-Berkeley). In addition, OSU-WIC working standards were analyzed by IRMS to ensure comparability across methods. Isotope values for both oxygen and hydrogen are reported as ratios relative to Viena Standard Mean Ocean Water (V-SMOW) in standard "delta" notation:

$$\delta\ [‰] = \left(\frac{R_{sample}}{R_{standard}} - 1\right) \times 1000, \tag{2}$$

where $R_{sample}$ and $R_{standard}$ are the ratios of $^{2}H/^{1}H$ (for $\delta^{2}H$) or $^{18}O/^{16}O$ (for $\delta^{18}O$) of the sample and V-SMOW, respectively.

## RESULTS

*Near-stream zone of influence experiment*

Tree felling was done within the time constraints of commercial logging activity at the site. Experiment A was impacted by an unexpected and inopportune cool, wet period that began the day we felled the trees and lasted several weeks. Results for this experiment were negated as a consequence of increased streamflow obscuring the diel fluctuations. As a result, we report only the results of Experiment B. Trees were felled for Experiment B on June 30, 2009 (Figure 4.4). Discharge measurements from June 27-29, 2009 were used to estimate the volume of “missing” water associated with the daily diel variations in streamflow and estimate the size of the near-stream “zone of influence” (following *Bond et al.* [2002], and as described in the methods). The daily mean volume of missing water during this period was 0.018 mm (or 464 l). This amount equated to calculated water consumption of transpiring vegetation covering a horizontal ground area of approximately 230 $m^2$ when using 2 mm $day^{-1}$ as an estimate of daily transpiration; consistent with estimates from the previous growing season. Since this area was less than the narrow riparian corridor, we conservatively delineated the area to be felled by estimating the position where the trees were likely to become disconnected from the stream and groundwater systems based on known groundwater table depths and the local topography. We determined that a zone of 5 m (h.d.) on either side of the stream and extending from the outlet up the channel 152 m, to the approximate perennial channel initiation point, would cover an adequate area of transpiring vegetation to shut down the diel fluctuation (estimated to transpire 0.12 mm $day^{-1}$—6.6 times the amount we estimated necessary to account for the fluctuations) and provide a sufficient distance from the stream so that the root systems of the trees occurring at or near the periphery of the experimental boundary would be out of reach of the groundwater. We felled 58 trees (43 Douglas-fir, 10 red alder, and 5 vine maple (*Acer circinatum*)) in the 1520 $m^2$ zone.

Diel fluctuations remained in the discharge record of Catchment B after the felling experiment (Figure 4.5). Flynn Creek, the control catchment, behaved similarly to Catchment B, exhibiting daily discharge fluctuations throughout the experimental period. The primary difference between the Catchment B and Flynn Creek records was that Flynn Creek had a steeper recession trend during the experiment. Daily transpiration rates were relatively uniform during the experiment as gauged by the consistent sap velocities measured in living trees within the catchment (Figure 4.5a).

We used the period from June 27-29 for the pre-treatment period and July 1-4 for the post-treatment period in our statistical analysis. To eliminate potential bias due to the different pre-treatment and post-treatment sample sizes (n=3 for pre-treatment and n=4 for post-treatment), we calculated post-treatment statistics for two different 3-day periods (*Post-treatment 1* is July 1-3, 2009 and *Post-treatment 2* is July 2-4, 2009). We did not detect change in the mean diel amplitude for either post-treatment period (p=0.56 and 0.79 for *Post-treatment 1* and *Post-treatment 2*, respectively; Table 2). The timing of the fluctuation was affected by the experimental felling as determined by a reduction in the mean treatment-control ratio of the lag between maximum transpiration and minimum stream discharge from 1.53 for the pre-treatment period to 0.92 for *Post-treatment 1* (p=0.07) and 0.73 for *Post-treatment 2* (p=0.01; deemed statistically significant). The change in the ratio was a result of a shorter lag at Catchment B in the post-treatment periods.

*Testing for influence beyond the near-stream zone*

Diel fluctuations persisted in the stream discharge record at Catchment B and in groundwater levels at Catchment A even after all trees had been cleared from the site (Figure 4.6). In addition to the trees, most understory vegetation was either removed or died during the harvest with the exception of areas along the ridge, well away from the stream channel (see Figure 4.6a). The discharge fluctuations for Catchment B followed the same general pattern as measured at the control catchment

(Flynn Creek). The Catchment A flume was compromised during the harvesting operation, so discharge measurements are not available for this period. However, water levels in Well-7HS and Well-8HS showed that groundwater in the bedrock underlying the Catchment A hillslope also continued to fluctuate on a diel cycle following the complete removal of trees.

*Isotope tracing to determine tree water source*

Antecedent wetness generally declined from June through early- to mid-August at which point it remained relatively constant through the end of September (Figure 4.7a). Several small summer precipitation events caused temporary, but small increases in API; these precipitation pulses were also evident in groundwater levels at the toeslope, streambank, and alluvial plain plots (Well-11SF, Well-14SF, and Well-16SF, respectively). The water table dynamics measured at our plots during the course of the growing season are shown in Figure 4.7b. Groundwater levels at the hillslope plot (Well-9HS) were relatively static during the period of record (this well was not installed until late-July); water levels were between 7.19 and 7.38 m below the ground's surface with most of the variation resulting from a drop in water level after the well was purged and sampled. Measured groundwater levels at the toeslope (Well-11SF) declined from 2.15 to 2.44 m below the ground's surface before going below the grade of the well in late-July. Groundwater at the streambank plot (Well-14SF) ranged from 0.24 to 0.66 m below the ground's surface, with minimum values occurring in August. On the alluvial plain (Well-16SF), groundwater was only 0.07 m below the ground's surface at the beginning of the growing season; the water level remained at that approximate depth until mid-July when it began decreasing until it reached a level approximately 0.4 m below the ground's surface in early-August. With the exception of several increases in response to precipitation events, it fluctuated around 0.4 m depth for the remainder of the measurement period. Both the streambank and alluvial plain groundwater showed strong diel fluctuations throughout the growing season.

The patterns of xylem water, soil water, surface water, and groundwater isotopes remained consistent for each landscape position across all sampling events. To facilitate interpretation, we report only the results from the 23 June 2009 and 24 September 2009 sampling events (Figure 4.8). Stream water and groundwater were tightly grouped and plotted on the global meteoric water line (GMWL) for both sampling periods. Soil water collected at 0.10 m and below at the hillslope and toeslope sites plotted just below the GMWL but within the range of precipitation variability (Figure 4.8a and c). The two soil water samples collected from the mineral soil surface and 0.05 m below the surface on 24 September 2009 plotted farther from the GMWL and outside of the range of precipitation variability. This feature is seen clearly in Figure 4.8c where the dark blue circles and square represent samples from the surface of the a-horizon. Soil water measured at the streambank and alluvial plain sites (Figure 4.8b and d) generally plotted along the GMWL, but with some separation from the groundwater and streamwater samples. The surface and 0.05 m soil samples collected on24 September 2009 from the streambank and alluvial plain were isotopically heavier, relative to the deeper soil water, but plotted within the range precipitation (Figure 4.8d). The mean $\delta^{2}H$ and $\delta^{18}O$ for the 12 mm of precipitation received during the 14 day period leading up to 24 September 2009 was -32 and -5, respectively.

Xylem water composition was grouped by species (Figure 4.8). In all sampling periods, Douglas-fir (hillslope and toeslope plots; Figure 4.8a and b) xylem water plotted within the vicinity of the soil water. Red alder xylem water (Figure 4.8b and d) plotted outside of the mixing-space of soil water, stream water, and groundwater for each sampling event. In all cases, our trees did not have an isotopic signature that suggested that they used streamwater or groundwater. As previously mentioned, the isotopic signature of xylem water remained consistent at each sampling date, and hence across the decreasing antecedent wetness conditions. The red alder anomaly is explored further in the discussion section.

## DISCUSSION

Low streamflow periods create stress-inducing conditions for lotic ecosystems [*Smakhtin*, 2001]. With decreasing stream discharge, aquatic habitat becomes marginal and, in some cases isolated or non-existent [*Gippel and Stewardson*, 1998], while the thermal buffering capacity regulating stream heating is minimized [*Brown*, 1969]. In the case of temperature-sensitive species, survival through the low-flow period is contingent on stream temperatures not deviating severely from the range tolerated by a particular species [*Groom et al.*, 2011]. From a societal perspective, public water supply and waste load allocation are both dependent on how much water is present in the stream [*Smakhtin*, 2001]. It is therefore important for resource managers to understand how changes in land use, particularly forest management, and climate change may affect stream low-flows. This is of utmost importance in regions that experience extended dry-seasons which coincide with the annual growing season, essentially creating a situation where ecosystem water demands are highest while water supply is lowest. Our study examined where low flow diel streamflow fluctuations were generated by testing the Bond et al. (2002) near-stream zone of influence hypothesis.

### *Near-stream zone of influence experiment*

We were unable to shut down the diel fluctuation in stream discharge with our near-stream “zone of influence” tree felling experiment. We therefore reject the *Bond et al.* [2002] near-stream “zone of influence” hypothesis at our site. Due to the extremely short pre-treatment and post-treatment periods, quantifying change, in a statistical sense, was very difficult. The timing and duration of our experiments were constrained by climate conditions and the larger industrial harvest operation that we operated within. While the landowners and contractors were extremely helpful in planning and facilitating our experiment, operational deadlines prevented the flexibility that would have been necessary to acquire the desired length of pre-treatment and post-treatment data records for each experiment during the prime diel fluctuation period (a function of stream discharge magnitude and climatic variables). So although we provide simple statistics to detect if the hydrograph responded to our treatment, we caution that inferences based on these statistics should be made with care.

*Bond et al.* [2002] ruled out the possibility that direct evaporation was causing the diel fluctuations observed at HJA-WS01. We also assessed the potential for direct evaporation from the water surface to cause the observed diel streamflow fluctuations, particularly following the experimental felling, by estimating potential evaporation with on-site measurements of temperature, wind speed, and relative humidity along with global radiation measurements from a nearby Ameriflux site (Marys River Fir Site, approximately 28 km northeast of Needle Branch, http://public.ornl.gov/ameriflux/Site_Info/siteInfo.cfm?KEYID=us.oregon_fir.01) using the Penman model [*Penman*, 1948]. To account for the daily cycles of the climate forcings used in the model, we calculated direct evaporation on hourly timesteps and then summed over the day to arrive at total daily direct evaporation. The mean estimate of direct evaporation from the Catchment B stream channel for the period of the experiment is 0.011 mm day$^{-1}$. This value is the same order of magnitude as the mean amount of water missing in the stream for the pre-treatment and post-treatment periods (0.012 mm day$^{-1}$ for both periods). However, this is an estimate of potential evaporation and we consider the effects of actual direct evaporation to be much less than the open-water-based calculation above might suggest because:

1) The stream was fully-shaded during the pre-treatment period. Therefore, the estimated direct evaporation is an over-exaggeration for this period. The stream continued to fluctuate during the post-treatment period of Experiment B in a similar manner to that observed during the pre-treatment period. This suggests that the mechanism driving the diel fluctuations did not change when the trees were felled (i.e. there was

not a switch from transpiration-driven to direct evaporation-driven diel streamflow fluctuations).

2) The stream remained fully- to partially-shaded during the post-treatment period as a result of the canopies of the felled trees lying over the top of the stream channel. This effect has been shown to cause stream temperatures to decrease following forest harvest in the headwater catchments of the Hinkle Creek Paired Watershed Study in the Oregon Cascades [*Kibler*, 2007]. The standing trees adjacent to the felled area also provided channel shade except during the periods of maximum solar angle.

3) The stream is deeply incised. Therefore the actual time the stream received direct radiation is limited to periods of high solar angle and subsequently is only a fraction of the duration of solar radiation used in the calculation. This is another factor leading us to believe the estimated direct evaporation is much greater than what actually occurs, even with no vegetation, standing or felled, available for shade.

Assuming direct evaporation is not responsible for the diel streamflow fluctuations, our manipulative experiment implies that the influence of vegetation on diel discharge fluctuations extends beyond the near-stream zone. Our initial alternative hypotheses to explain this result included the possibility of water extraction by roots of upslope vegetation that extended downslope into the near-stream zone. Douglas-fir growing on steep slopes have been shown to have proportionally longer downslope roots [*McMinn*, 1963]. Therefore, it is possible that upslope trees located outside of our experimental felling area were rooted far enough into the near-stream zone to create the diel fluctuation (we expected that such trees would have roots within the zone, but the design assumed that these roots would not be influential in a diel fluctuation sense). Another alternative hypothesis was that the vegetation located in the convergent zone immediately upslope from the stream head controlled the diel fluctuations. Both of these alternative hypotheses would require that the diel fluctuation be eliminated following complete vegetation removal.

*Complete vegetation removal: Diel influences beyond the near-stream zone?*

Complete vegetation removal in both Catchment A and Catchment B did not shut down the diel fluctuations in streamflow or groundwater. This was an unexpected result and prompted further scrutiny of our data. Our quality control measures, that included manual stream stage and groundwater level measurements as well as instrument calibration checks, showed no sign of compromised data. However, other studies have questioned whether diel fluctuations they measured in streamflow were 'real'. *Cuevas et al.* [2010] tested their instrumentation and found that the barometric correction for their non-vented pressure transducers, which were the same transducer type and model that we used to monitor groundwater in Catchment A (used in all HS and FP series wells), had a temperature dependency that created diel fluctuations in the data record. Pressure measurements are temperature-dependent and non-vented pressure transducers require a correction for barometric pressure (typically using a senor of the same model deployed in the atmosphere). When the ambient temperature of the barometric pressure sensor and water-level sensor are significantly different, they provide under- or over-estimates, depending on the direction of the temperature differential, of the true water depth following the barometric correction procedure. Our barometric compensation sensor was deployed in the upper casing of one of our wells during the 2009 growing season. Although this location provided some thermal buffering, the temperature differential between the barometric and water level sensors was large enough to warrant further quality assurance tests of our data.

We conducted an experiment to test if the diel groundwater fluctuations we measured in the cleared catchment were real. We installed a section of PVC pipe, matching the material used for our other wells and capped and sealed at its base, to a depth of 2 m near Well-7HS and Well-8HS. The pipe was filled with water and a water level sensor of the type in question was deployed during the 2011 growing season. Figure 4.9 shows the results relative to Well-8HS. The water level record from the sealed pipe (vented to the atmosphere) showed some sign of the behavior described by *Cuevas et al.* [2010] with small increases in water level (on the order of 0.0025 m) occurring at mid-day when the temperature differential was greatest. However, this fluctuation is smaller and not of the same timing as the fluctuations observed at Well-7HS and Well-8HS immediately after the forest was completely removed in Catchment A during the 2009 growing season.

Based on these findings, we believe that the fluctuations measured in Well-7HS and Well-8HS were actual changes in water level and not an artifact of temperature differentials between our sensors. Although *Cuevas et al.* [2010] identified this temperature-dependent measurement artifact, they found that under normal field condition it only served to slightly enhance true fluctuations in stream stage (by approximately 17%). Overall we reject both alternative hypotheses posed in the previous section because of the continued fluctuations measured in both streamwater and groundwater after all the vegetation within the catchment was eliminated. Our results therefore suggest that we cannot limit the generation source area of the diel fluctuations to the immediate catchment area.

*Isotope tracing to determine tree water source*

We were unable to link, isotopically, the water sources trees were utilizing to water that was contributing to streamflow. Streamflow appeared to be sourced solely from groundwater during the growing season based on the close grouping of groundwater and streamwater isotope values. The lack of soil water alignment with stream water and groundwater along the GMWL suggests that the soil water was influenced by recent precipitation, whereas the stream and groundwater integrate a longer history of inputs (mean transit time of soil water is 3-4 months while that of stream water is 4 to 10 years, *Hale et al.* [2011] ).Our attempt to distinguish tree source waters suggested that Douglas-fir occupying hillslope and toeslope positions use a mixture of water from different soil depths (Figure 4.8a and c). They do not appear to use any groundwater or stream water. This finding is not unexpected for the hillslope plot given that the depth to groundwater is approximately 7 m below the ground surface and the stream is approximately 40 m (slope distance) from the plot. However, the toeslope sampling location was within reasonable reach of both the groundwater and stream based on known Douglas-fir rooting behavior [*McMinn*, 1963]. The finding that our trees only use soil water is in agreement with recent work by *Brooks et al.* [2010] and *Goldsmith et al.* [2011] showing that there is a separation between the water trees utilize for transpiration and the more mobile water that contributes to groundwater recharge and streamflow. The red alder trees in our streambank and alluvial plain plots did not appear to be using either soil water or streamwater.

Since no trees were found to be using water linked to streamflow, our question of whether trees switch water sources under declining antecedent moisture conditions does not apply as originally posed. However, the consistency in xylem water isotopic signatures throughout the growing season indicates that their water sources are stable in time and space. The *Bond et al.* [2002] hypothesis posits that trees "disconnected" from the hydrologic system during the course of the seasonal drought as a result of the water table declining beyond the reach of the roots systems. This could be plausible at our toeslope plot where the groundwater decreased below the base level of our well (at 2.4 m depth) mid-way through the growing season. However, the trees at this plot showed no sign of using groundwater. In the streambank and alluvial plain plots, where we expected the trees to be utilizing groundwater, the water table elevation did not decrease significantly; groundwater levels fluctuated by a maximum of 0.4 m at both plots. Even at their minimum elevation, the groundwater at these two plots was well within the typical rooting depth of these trees [*Harrington et al.*, 1994], and as observed during our auguring to collect soil samples (roots common to at least 1 m).

The red alder xylem water anomaly, whereby the trees plotted outside of the mixing space of all potential source waters that we measured, remains vexing. The trees in the streambank plot had live roots extending into the stream and streambed. The depth to groundwater at the streambank and alluvial plain plots was relatively shallow for the duration of the growing season (less than 0.7 m). We therefore expected that these trees would be at least partially using these water sources. However, our red alder trees did not appear to be using any of the sources we measured, unlike the findings of *Dawson and Ehleringer* [1991], where streamside trees were found not to be using streamwater but actually a deeper groundwater source. It is possible, yet highly unlikely, that we missed a water source even though we sampled soil water at multiple depths, shallow groundwater (1 m), deeper groundwater (5-7 m), and stream water.

Another possible explanation for this puzzling finding is that there could be fractionation occurring at the soil-root interface or within the plant that we have not taken into account. While fractionation associated with water uptake by plants is relatively rare, hydrogen fractionation at the root has been observed in some halophytes and xerophytes [*Ellsworth and Williams*, 2007; *Lin and Sternberg*, 1993; *Sternberg and Swart*, 1987]. Although the fractionation mechanism is unknown, it is assumed that it is a result of adaptations at the cellular-level which allow the plants to survive in saline environments [*Ellsworth and Williams*, 2007]. While red alder do not inhabit saline environs, they do have a symbiotic relationship with the nitrogen-fixing *Frankia* species of actinomycetes [*Molina et al.*, 1994]. Colonization of the root cells of red alder by the actinomycetes is ubiquitous and results in the formation of root nodules which are dense with aerenchyma cells that allow gas exchange with the atmosphere. Although review of the literature failed to produce a defensible fractionation mechanism associated with the nitrogen-fixing processes within the root or the presence of aerenchyma cells, it is possible these features could cause fractionation which would explain the isotopic composition of the red alder xylem water. Although water isotopes can be useful tools for constraining the water sources utilized by plants, our results show the clear need for further work in determining the potential for fractionation upon uptake or within the plant on a species-by-species basis.

*A mechanism for diel streamflow fluctuations?*

So what might generate diel fluctuations in streamflow in a completely deforested catchment? Assuming direct evaporation from the channel plays a minimal role in the persistent diel streamflow and groundwater fluctuations, we hypothesize that diel signals generated by vegetation in the lower, forested portion of the catchment are transported via saturated bedrock fractures to the groundwater system in the upper catchment. We suggest that the observed change in water level is then a result of changes in hydraulic gradient being transferred through the fracture system—not a result of direct water flux, which, based on reasonable hydraulic conductivity values, could not occur on the timescale necessary to create the diel fluctuations in the upper catchment.

Our finding of diel fluctuations being transported over long distances via bedrock fractures supports the *Wondzell et al.* [2010] supposition that the whole catchment should be considered when assessing the controls on diel streamflow fluctuations. Although we are unaware of any other studies documenting the propagation of diel fluctuations through fracture networks, other studies do provide evidence that the generation of diel fluctuations is not limited to the near-stream area. Diel streamflow fluctuations were only partially damped when *Dunford and Fletcher* [1947] removed all vegetation within 4.6 m elevation above the stream channel, equivalent to 12 percent of the total catchment area, in a small stream at the Coweeta Hydrologic Laboratory in the southern Appalachian Mountains (near Otto, NC). This, as well as the finding by *Barnard et al.* [2010] that diel fluctuations were present in the trenched outflow of the HJA-WS10 hillslope, provides substantial evidence that diel fluctuations are generated in places other than the near-stream zone. Even when the location of the signal generation area is clear, such as in semi-arid systems where vegetation only occurs along the riparian corridor, many studies have shown that the speed by which the signal is propagated to the stream varies depending on aquifer properties specific to a particular flowpath [*Butler et al.*, 2007; *Gribovszki et al.*, 2008; *Szilagyi et al.*, 2008]. Based on our findings, as well as the other examples from the literature, diel fluctuations are, in most cases, generated, propagated, and attenuated at varying rates and magnitudes throughout catchment. Thus, the fluctuation at any given point within the catchment or stream network is an unknown combination of the myriad signals created within the catchment.

*Diel fluctuations as a hydrological red herring?*

The study of diel fluctuations in stream discharge is predicated on the potential information content that the diel signal could contain relating to how water is stored and transported within the catchment and how vegetation interacts with these processes [*Bond et al.*, 2002; *Gribovszki et al.*, 2010; *Wondzell et al.*, 2007]. It is now clear that the generation of these diel fluctuations is not a simple impulse-response mechanism as previously thought [*Wondzell et al.*, 2010]. Although modern instrumentation is sensitive enough to measure the changes in water elevation that occur with diel fluctuations, the "missing" water volumes resulting from the diel suppression of discharge are within the uncertainty range of our ability to accurately gauge discharge (Table 3). Thus, as we must push the bounds of our sensing capabilities to measure a signal known to be an amalgam of many interacting signals generated throughout the catchment, we should ask ourselves what is the value of investing further research efforts into the cause of understanding hydrologic and ecohydrologic mechanisms by way of diel fluctuations in stream discharge. Alas, are these fluctuations just a hydrological red herring [*Beven and Westerberg*, 2011]?

## CONCLUSIONS

Our study tested the *Bond et al.* [2002] hypothesis that a near-stream "zone of influence" was the source area for generating diel streamflow fluctuations using a vegetation removal experiment and isotope tracing to better understand how and where within a catchment forest vegetation influences streamflow during the growing season. Following tree felling in a "zone of influence" delineated following the methods of *Bond et al.* [2002], diel streamflow fluctuations continued, leading us to reject the "zone of influence" hypothesis. Diel fluctuations in streamflow and groundwater persisted after all forest vegetation had been removed from the catchment (i.e. completely clear-cut); indicating the source-area for the fluctuations extended even beyond the immediate catchment boundaries. We hypothesized that fluctuations generated in the still-forested lower section of the catchment were propagated to the upper, deforested catchment through groundwater confined in bedrock fractures. Our isotope tracing results showed that streamwater and groundwater were distinctly different than soil water. Douglas-fir trees located at hillslope and toeslope positions appeared to only utilize soil water for transpiration and, therefore, were not directly sourcing water contributing to streamflow. Red alder trees located on the streambank and the alluvial plain where the water table was closer to the surface did not appear to be using any of the water sources we measured. Further work is necessary to determine if the trees are using an alternative water source or if the isotopic signature of the xylem water is altered by physiological processes.

As a result of this new understanding, caution should be made when using diel fluctuations to estimate evapotranspiration. Those applying the practice should take care to make sure that the fluctuations used in their calculations are generated by local sources and not influenced by adjacent or distant catchment components.

## REFERENCES

Asbjornsen, H., et al. (2011), Ecohydrological advances and applications in plant–water relations research: a review, *Journal of Plant Ecology*, *4*(1-2), 3-22.

Barnard, H. R., C. B. Graham, W. J. Van Verseveld, J. R. Brooks, B. J. Bond, and J. J. McDonnell (2010), Mechanistic assessment of hillslope transpiration controls of diel subsurface flow: a steady-state irrigation approach, *Ecohydrology*, *3*(2), 133-142.

Beven, K., and I. Westerberg (2011), On red herrings and real herrings: disinformation and information in hydrological inference, *Hydrol. Processes*, n/a-n/a.

Bond, B. J., J. A. Jones, G. Moore, N. Phillips, D. Post, and J. J. McDonnell (2002), The zone of vegetation influence on baseflow revealed by diel patterns of streamflow and vegetation water use in a headwater basin, *Hydrol. Processes*, *16*(8), 1671-1677.

Brooks, J. R., H. R. Barnard, R. Coulombe, and J. J. McDonnell (2010), Ecohydrologic separation of water between trees and streams in a Mediterranean climate, *Nature Geoscience*, *3*(2), 100-104.

Brown, G. W. (1969), Predicting temperatures of small streams, *Water Resour. Res.*, *5*(1), 68-75.

Butler, J. J., Jr., G. J. Kluitenberg, D. O. Whittemore, S. P. Loheide, II, W. Jin, M. A. Billinger, and X. Zhan (2007), A field investigation of phreatophyte-induced fluctuations in the water table, *Water Resour. Res.*, *43*(2), W02404.

Constantz, J. (1998), Interaction between stream temperature, streamflow, and groundwater exchanges in alpine streams., *Water Resour. Res.*, *34*(7), 1609-1615.

Corliss, J. F. (1973), Soil Survey, Alsea Area, Oregon, *U.S. Soil Conservation Service*, 82 pp.

Cuevas, J. G., M. Calvo, C. Little, M. Pino, and P. Dassori (2010), ARE DIURNAL FLUCTUATIONS IN STREAM FLOW REAL?, *Journal of Hydrology and Hydromechanics*, *58*(3), 149-162.

Dawson, T. E., and J. R. Ehleringer (1991), Streamside trees that do not use stream water, *Nature*, *350*(6316), 335-337.

Descroix, L., J. F. Nouvelot, and M. Vauclin (2002), Evaluation of an antecedent precipitation index to model runoff yield in the western Sierra Madre (North-west Mexico), *J. Hydrol.*, *263*(1-4), 114-130.

Dunford, E. G., and P. W. Fletcher (1947), Effect of removal of stream-bank vegetation upon water yield, *Eos Trans. AGU*, *28*, 105-110.

Ehleringer, J. R., and C. B. Osmond (1989), Stable Isotopes, in *Plant Physiological Ecology Field Methods and Instrumentation*, edited by R. W. Pearcy, J. R. Ehleringer, H. A. Mooney and P. W. Rundel, pp. 281-300, Chapman and Hall, Ltd., London.

Ehleringer, J. R., and T. E. Dawson (1992), WATER-UPTAKE BY PLANTS - PERSPECTIVES FROM STABLE ISOTOPE COMPOSITION, *Plant Cell and Environment*, *15*(9), 1073-1082.

Ellsworth, P. Z., and D. G. Williams (2007), Hydrogen isotope fractionation during water uptake by woody xerophytes, *Plant and Soil*, *291*(1-2), 93-107.

Gat, J., and Y. Tzur (1967), Modification of the isotopic composition of rainwater by processes which occur before groundwater recharge, in *Proceedings of the Symposium on Isotopes in Hydrology*, edited, pp. 49-60, International Atomic Energy Agency, Vienna, Austria, 14-18 November 1966.

Gippel, C. J., and M. J. Stewardson (1998), Use of wetted perimeter in defining minimum environmental flows., *Regulated Rivers: Research and Management*, *14*, 53-67.

Goldsmith, G. R., L. E. Muñoz-Villers, F. Holwerda, J. J. McDonnell, H. Asbjornsen, and T. E. Dawson (2011), Stable isotopes reveal linkages among ecohydrological processes in a seasonally dry tropical montane cloud forest, *Ecohydrology*, n/a-n/a.

Granier, A. (1987), Evaluation of transpiration in a Douglas-fir stand by means of sap flow measurements, *Tree Physiology*, *3*, 309-320.

Gribovszki, Z., J. Szilagyi, and P. Kalicz (2010), Diurnal fluctuations in shallow groundwater levels and streamflow rates and their interpretation - A review, *J. Hydrol.*, *385*(1-4), 371-383.

Gribovszki, Z., P. Kalicz, J. Szilagyi, and M. Kucsara (2008), Riparian zone evapotranspiration estimation from diurnal groundwater level fluctuations, *J. Hydrol.*, *349*(1-2), 6-17.

Groom, J. D., L. Dent, and L. J. Madsen (2011), Stream temperature change detection for state and private forests in the Oregon Coast Range, *Water Resour. Res.*, *47*(1), W01501.

Hale, V. C., J. J. McDonnell, M. K. Stewart, D. K. Solomon, G. G. Ice, and R. T. Pack (this issue), Similar catchment form and rainfall-runoff responses hide radically different plumbing: (2) The role of catchment storage controls, *to be submitted to Water Resources Research*.

Harr, R. D. (1976), Forest practices and streamflow in western Oregon. Gen. Tech. Rep. PNW-49*Rep.*, 18 pp, U.S. Department of Agriculture, Forest Service, Pacific Northwest Forest and Range Experiment Station, Portland, OR.

Harrington, C. A., J. C. Zasada, and E. A. Allen (1994), Biology of red alder (*Alnus rubra* Bong.), in *The Biology and Management of Red Alder*, edited by D. E. Hibbs, D. S. DeBell and R. F. Tarrant, pp. 3-22, Oregon State University Press, Corvallis, OR

Harris, D. D. (1977), Hydrologic changes after logging in two small Oregon coastal watersheds, *United States Geological Survey*, *Water-Supply Paper 2037*.

Hewlett, J. D. (1982), *Principles of Forest Hydrology*, 183 pp., University of Georgia Press, Athens, GA.

Jones, J. A., and D. A. Post (2004), Seasonal and successional streamflow response to forest cutting and regrowth in the northwest and eastern United States, *Water Resour. Res.*, *40*(W05203), doi:10.1029/2003WR002952.

Kibler, K. M. (2007), The influence of contemporary forest harvesting on summer stream temperatures in headwater streams of Hinkle Creek, Oregon, 115 pp, MS Thesis, Oregon State University, Corvallis, OR.

Lin, G. H., and L. D. L. Sternberg (1993), EFFECTS OF SALINITY FLUCTUATION ON PHOTOSYNTHETIC GAS-EXCHANGE AND PLANT-GROWTH OF THE RED MANGROVE (RHIZOPHORA-MANGLE L), *J. Exp. Bot.*, *44*(258), 9-16.

Lundquist, J. D., and D. R. Cayan (2002), Seasonal and spatial patterns in diurnal cycles in streamflow in the western United States, *Journal of Hydrometeorology*, *3*(5), 591-603.

McMinn, R. G. (1963), CHARACTERISTICS OF DOUGLAS-FIR ROOT SYSTEMS, *Canadian Journal of Botany*, *41*(1), 105-122.

Meinzer, O. E. (1927), Plants as indicators of ground water, *U.S. Geological Survey Water-Supply Paper 577*, 95.

Molina, R., D. Myrold, and C. Y. Li (1994), Root symbioses of red alder: Technological opportunities for enhanced regeneration and soil improvement, in *The Biology and Management of Red Alder*, edited by D. E. Hibbs, D. S. DeBell and R. F. Tarrant, pp. 23-46, Oregon State University Press, Corvallis, OR

Moore, G. W., B. J. Bond, J. A. Jones, N. Phillips, and F. C. Meinzer (2004), Structural and compositional controls on transpiration in 40-and 450-year-old riparian forests in western Oregon, USA, *Tree Physiology*, *24*(5), 481-491.

Penman, H. L. (1948), Natural evaporation from open water, bare soil, and grass, *Proceedings of the Royal Society of London*, *A193*, 120-146.

Raats, P. A. C. (2007), Uptake of water from soils by plant roots, *Transport in Porous Media*, *68*(1), 5-28.

Rowe, P. B. (1963), Streamflow increases after removing woodland-riparian vegetation from a southern California watershed, *Journal of Forestry*, *365-370*.

Smakhtin, V. U. (2001), Low flow hydrology: a review, *J. Hydrol.*, *240*, 3-4.

Sternberg, L. d. S. L., and P. K. Swart (1987), Utilization of Freshwater and Ocean Water by Coastal Plants of Southern Florida, *Ecology*, *68*(6), 1898-1905.

Steudle, E. (2000), Water uptake by plant roots: an integration of views, *Plant and Soil*, *226*(1), 45-56.

Szilagyi, J., Z. Gribovszki, P. Kalicz, and M. Kucsara (2008), On diurnal riparian zone groundwater-level and streamflow fluctuations, *J. Hydrol.*, *349*(1-2), 1-5.

Thorson, T. D., S.A. Bryce, D.A. Lammers, A.J. Woods, J.M. Omernik, J. Kagan, D.E. Pater, and J.A. Comstock (2003), Ecoregions of Oregon, *U.S. Geological Survey, Reston, VA.*

Troxell, M. C. (1936), The diurnal fluctuation in the groundwater and flow of the Santa Ana River and its meaning, *Trans. AGU*, *17*, 496-504.

West, A. G., G. R. Goldsmith, P. D. Brooks, and T. E. Dawson (2010), Discrepancies between isotope ratio infrared spectroscopy and isotope ratio mass spectrometry for the stable isotope analysis of plant and soil waters, *Rapid Communications in Mass Spectrometry*, *24*(14), 1948-1954.

White, W. N. (1932), A method of estimating ground-water supplies based on discharge by plants and evaporation from soil: Results of investigations in

Escalante Valley, Utah, *U.S. Geological Survey Water-supply Paper 659-A*, 105.

Wicht, C. L. (1941), Diurnal fluctuations in Jonkershoek streams due to evaporation and transpiration, *Journal of the South African Forestry Association*, *7*, 34-49.

Wondzell, S. M. (2006), Effect of morphology and discharge on hyporheic exchange flows in two small streams in the Cascade Mountains of Oregon, USA, *Hydrol. Processes*, *20*(2), 267-287.

Wondzell, S. M., M. N. Gooseff, and B. L. McGlynn (2007), Flow velocity and the hydrologic behavior of streams during baseflow, *Geophys. Res. Lett.*, *34*(24).

Wondzell, S. M., M. N. Gooseff, and B. L. McGlynn (2010), An analysis of alternative conceptual models relating hyporheic exchange flow to diel fluctuations in discharge during baseflow recession, *Hydrol. Processes*, *24*(6), 686-694.

Table 4.1. Mean *treatment-control* ratios (Catchment B: Flynn Creek) of daily fluctuation amplitude and daily lag in discharge suppression (relative to sap velocity) for near-stream felling experiment. Numbers in parentheses indicate p-values for a two-sample t-test for a difference from the pre-treatment mean. The null hypothesis of no difference in mean is rejected for $p<0.05$ (values in bold).

| **Period** | **Mean Amplitude Ratio** | **Mean Lag Ratio** |
|---|---|---|
| Pre-treatment | 0.41 | 1.53 |
| Post-treatment 1 | 0.39 (0.56) | 0.92 (0.07) |
| Post-treatment 2 | 0.40 (0.79) | 0.73 **(0.01)** |

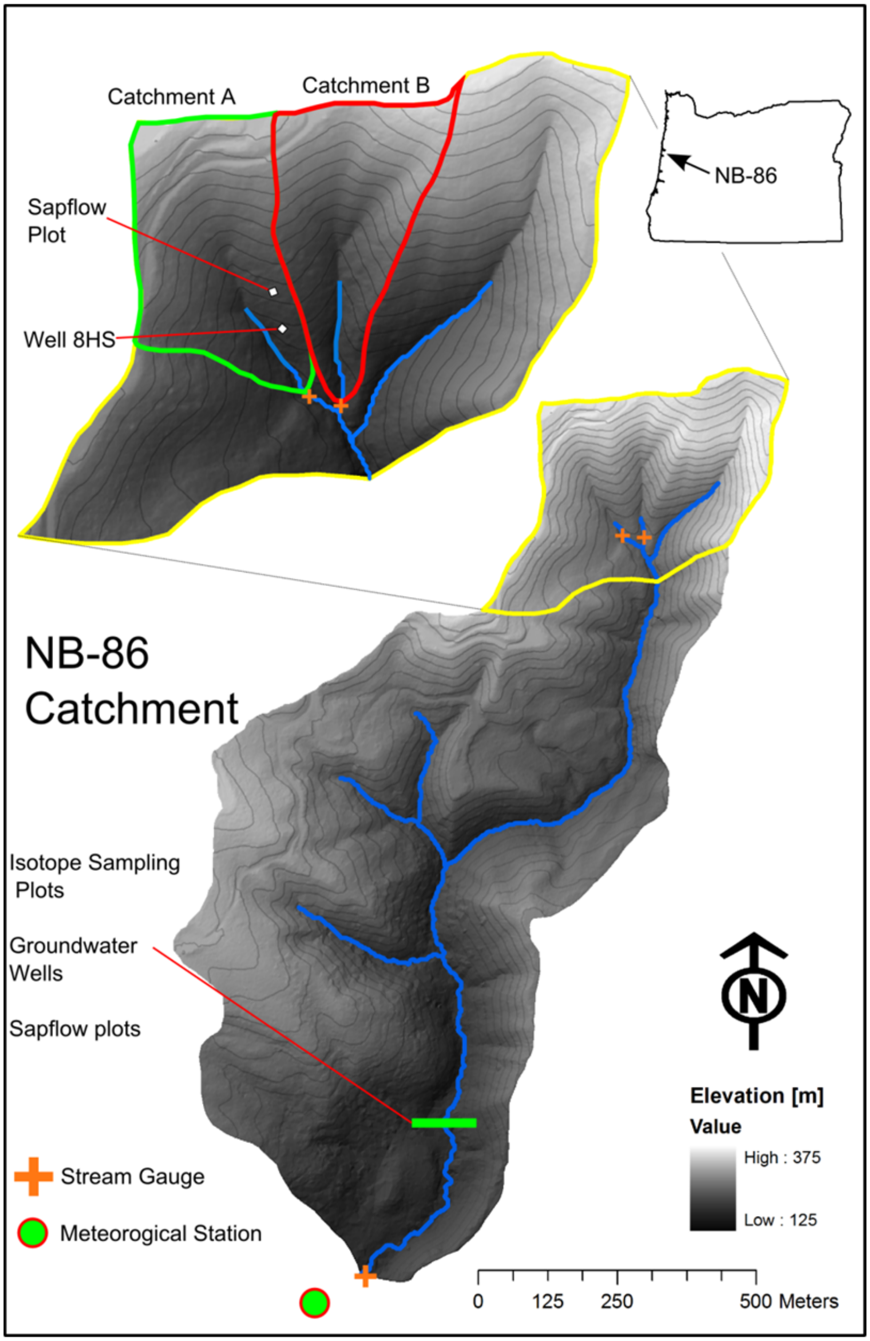


Figure 4.1. Study site map. Contour intervals are 10 m.

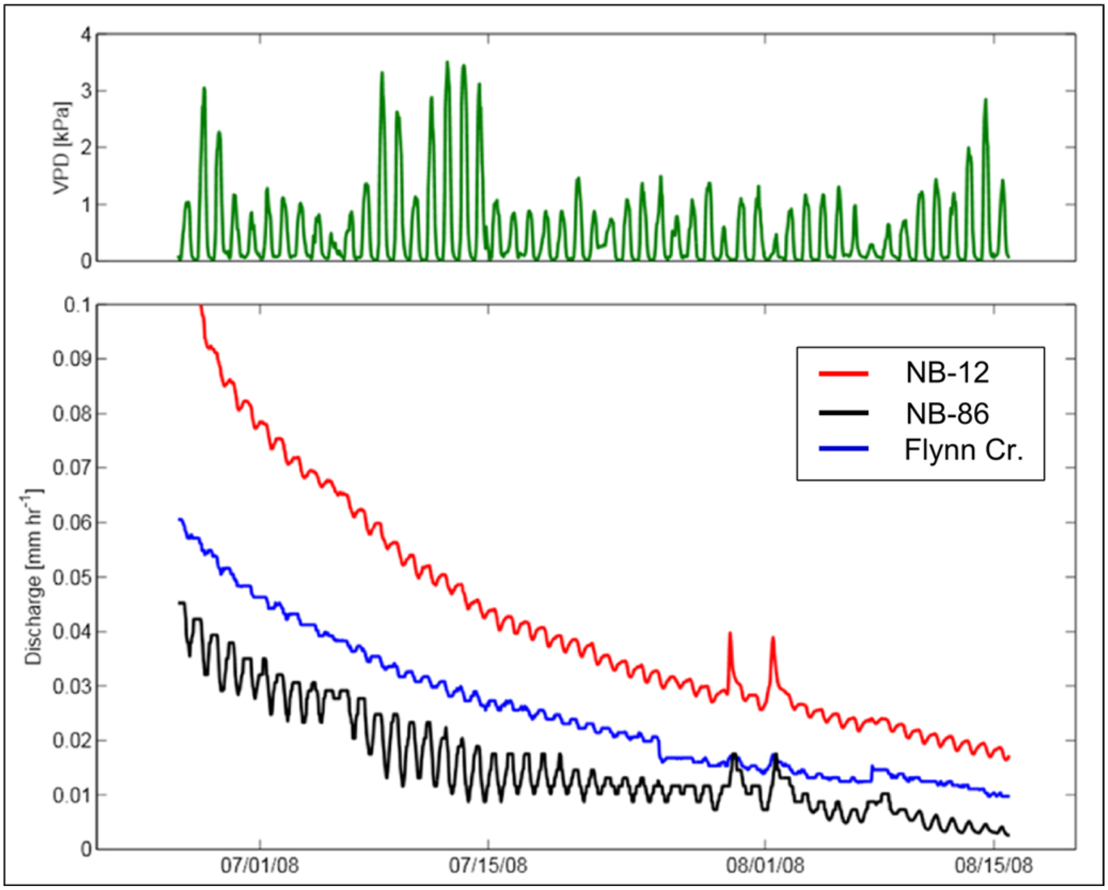


Figure 4.2. Example of diel streamflow suppression observed at two locations in our nested experimental catchment. The solid black line is discharge measured the smallest scale (2.5 ha catchment; 1$^{st}$-order; simple geomorphic character) and the dashed gray line is the largest scale (70 ha catchment; 2$^{nd}$-order; geomorphology includes alluvial plain and associated alluvial aquifer).

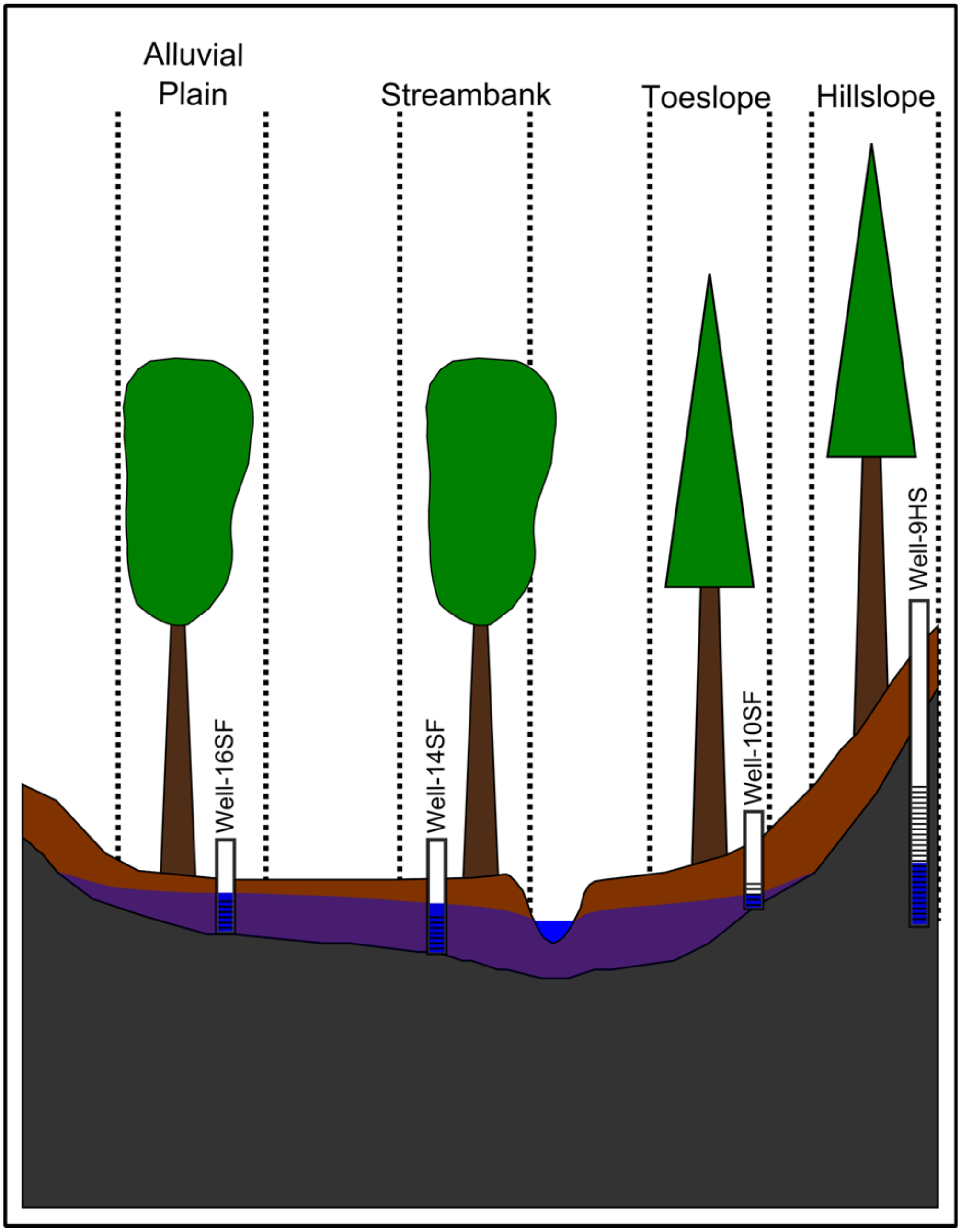


Figure 4.3. Diagram depicting the locations of sapflow and isotope sampling plots in the lower, non-harvested unit of the Needle Branch catchment (not to scale).

Figure 4.4. Photographs of experimentally felled near-stream “strips” as seen from above (top) and from the upstream extent of felled area looking downstream (bottom).

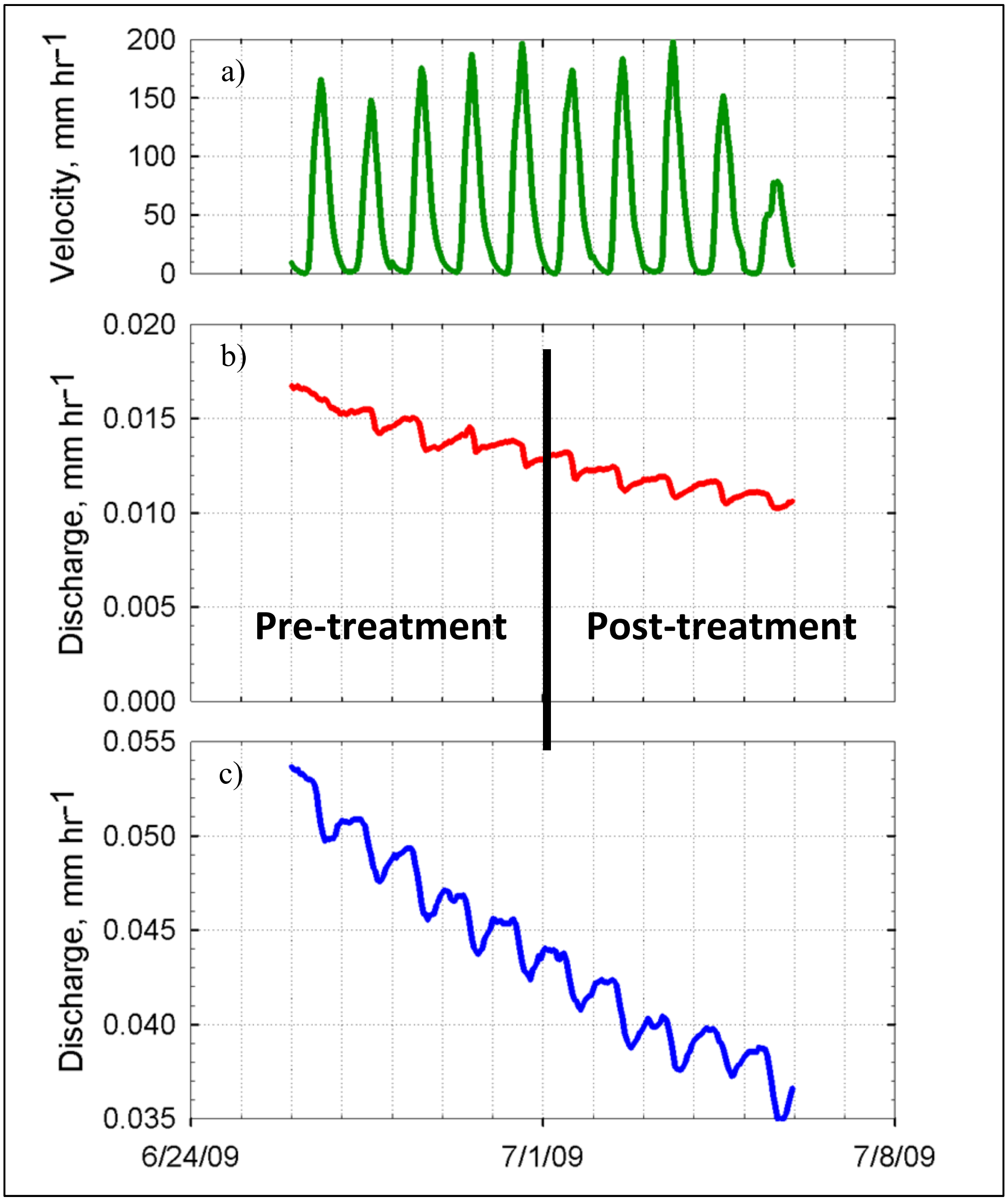


Figure 4.5. Mean sapflow velocity (a) and discharge measured for the 8-day period from June 26, 2009 to July 4, 2009 for the treatment (b) and control (c) streams. The experimental felling occurred on the morning of June 30, 2009 as indicated by the vertical line.

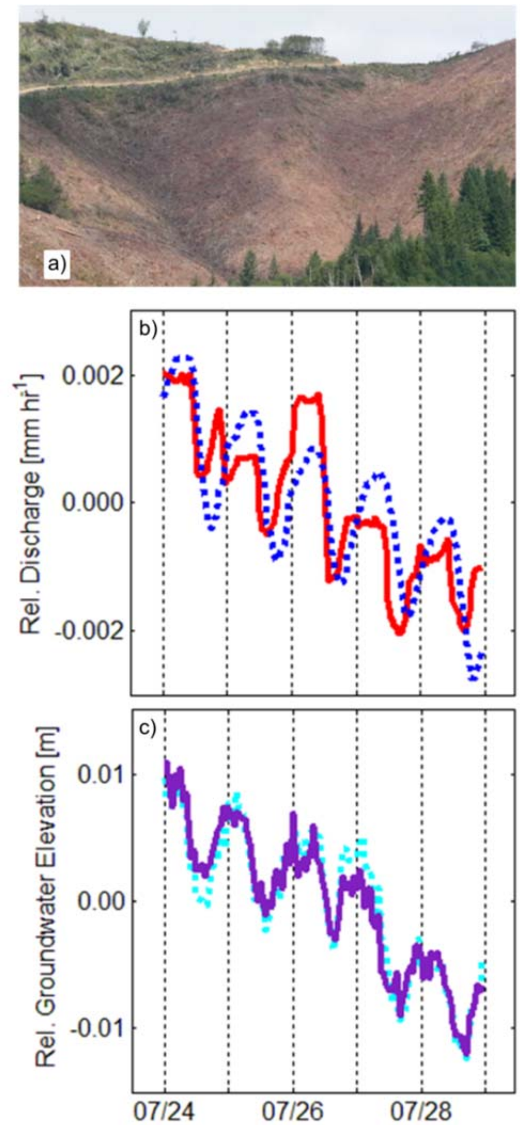


Figure 4.6. Photograph of the treatment catchment on July 29, 2009 (a), relative (mean subtracted) discharge for Catchment B (treatment; solid red line) and Flynn Creek (control; dotted blue line) (b), and relative ground water elevations measured at Well-7HS and Well-8HS within Catchment A (c).

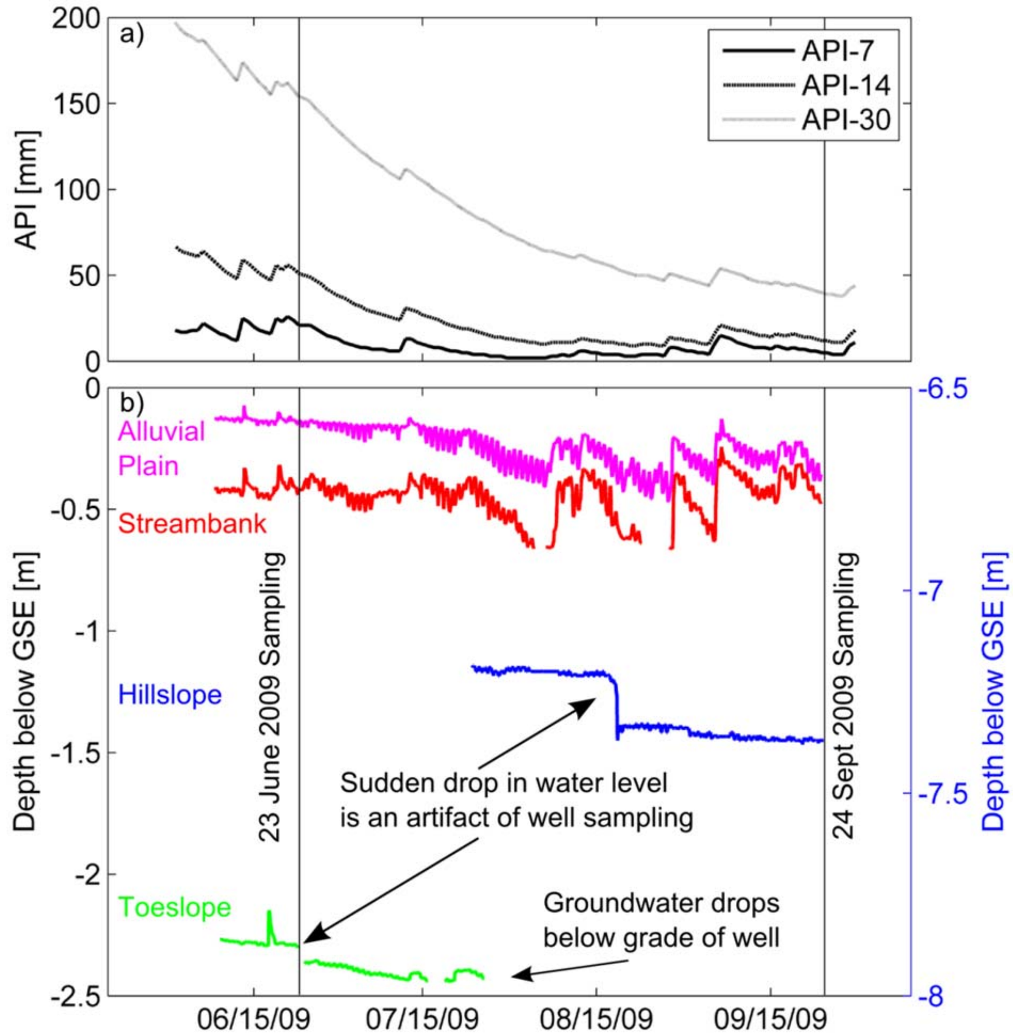


Figure 4.7. Antecedent precipitation index (API) and groundwater levels measured during the 2009 growing season. Sampling dates shown in Figure 4.8 are marked.

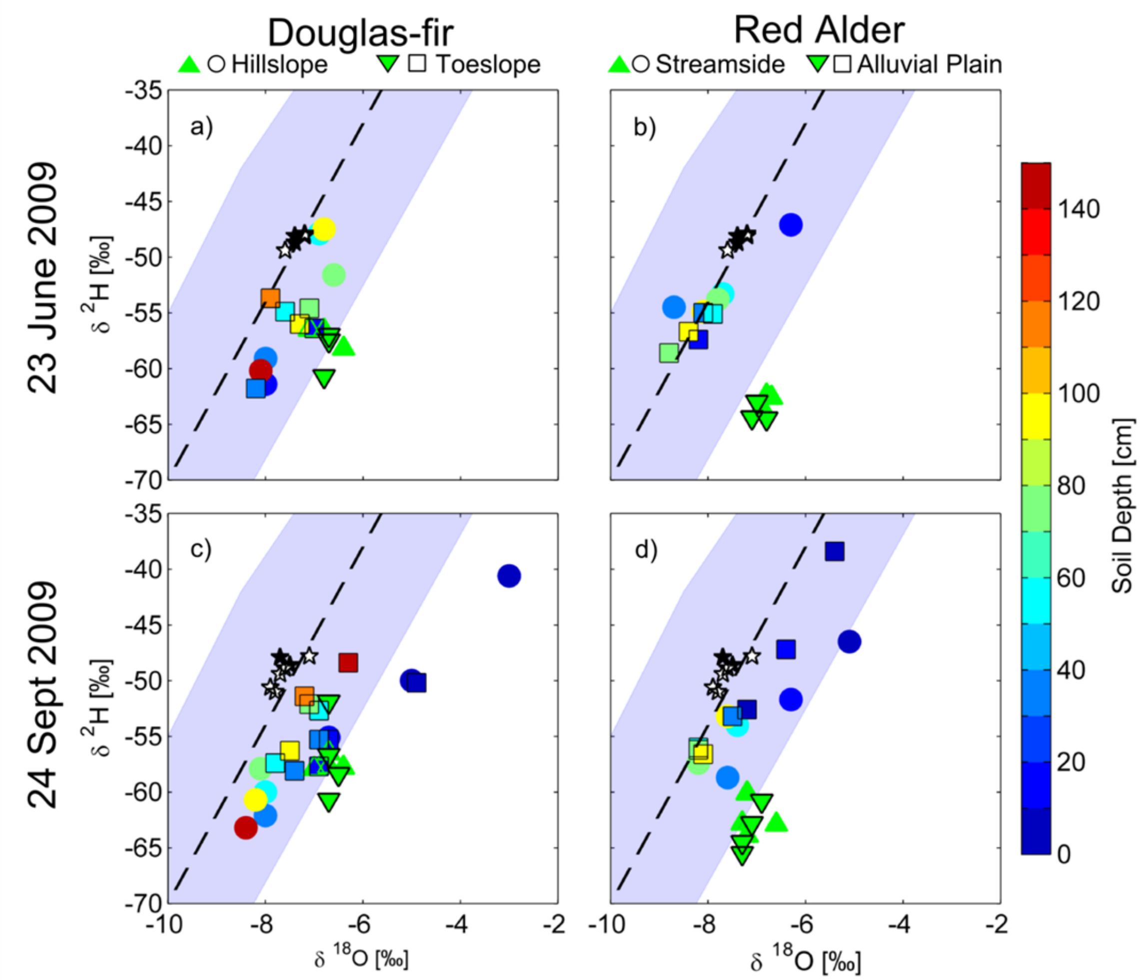


Figure 4.8. Groundwater (open stars), surface water (closed stars), soil water (squares and circles), and xylem water (triangles) stable isotope values for $\delta^2$H and $\delta^{18}$O. The shaded region indicates the range of precipitation isotope values. The global meteoric water line ($\delta^2H=8*\delta^{18}O+10$; dashed line) is plotted for reference.

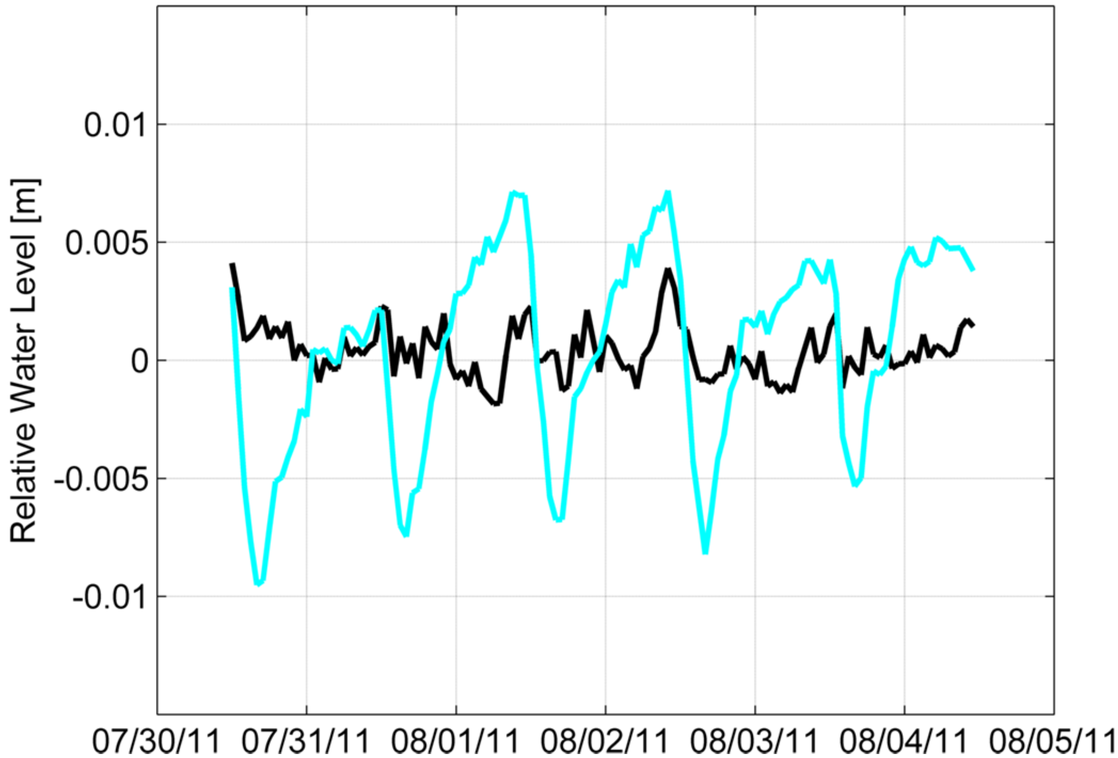


Figure 4.9. Relative water level measured in Well-8HS (blue line) and the capped pipe (black line) installed nearby.

# Conclusions

**CONCLUDING REMARKS**

Despite a rich history of forest hydrology research, critical questions remain regarding how forest management practices affect catchment hydrology and water quality [Andréassian, 2004]. The paired-catchment approach traditionally used by forest hydrologists focuses on measuring the hydrologic response to various forest management practices rather than the mechanisms and interactions that cause the response. This leads to a lack of process-level understanding and presents a problem when trying to use the findings from paired-catchment studies to predict the response of catchments in other settings (or at different scales) to the same forest management practices. In this dissertation, we changed the question from "What is the response to land-use change?" to "What causes the response to land-use change?", integrating a process-based hydrological investigation with an ongoing paired-catchment study to address fundamental gaps in understanding how forests interact with the hydrological system at multiple catchment scales.

Scaling process-understanding from hillslopes and small catchments to larger, management-relevant scales is one of the greatest challenges in hydrology; yet scaling is an essential component of predicting the hydrological effects of forest management. In Chapter 2 we investigated the role of geology in controlling catchment function across multiple scales. We used streamwater mean transit time (MTT), which is a useful proxy for catchment function as it integrates how water is stored and travels through the subsurface, within a catchment intercomparison framework to elucidate the geologic controls on hydrological scaling relationships. We compared MTT scaling relationships developed by *McGuire et al.* [2005] at volcanic catchments in the Western Cascades Range of Oregon, USA to those from our sedimentary geology research catchments in the Oregon Coast Range. We found fundamentally different scaling relationships between the two sites, despite striking similarities in catchment form, cover, and runoff-response.

In Chapter 3 we used a bottom-up approach to explore the process-reasons for the contrasting MTT scaling behavior we observed in the intercomparison study reported in Chapter 2. We combined hydrometric and tracer methods to investigate how bedrock permeability influenced catchment storage, MTTs, and MTT scaling relationships. We found that the permeable sedimentary bedrock in our Oregon Coast Range catchments led to the development of distinct zones of storage with different characteristic ages, whereas the tight geology of the Western Cascades catchments limited subsurface storage to the soil profile. Our work suggests that permeability contrasts in the subsurface represent perhaps the most basic control on how water is stored within the catchment and therefore is possibly one of the best direct predictors of streamwater MTT. These results suggest that although it may be tempting to conclude that catchments with similar outward appearances and hydrologic behavior function similarly below the surface, this is not always the case. Differences in catchment storage, and hence function, as controlled by bedrock permeability, could play a large role in controlling the hydrological response to environmental change, including forest harvesting.

Chapter 4 examined how and where forest trees interacted with subsurface storage during the growing season using a manipulative experiment. We tested the null hypothesis that near-stream trees alone influenced daily fluctuations in streamflow as posed by *Bond et al.* [2002]. We combined our manipulative experiment with isotope tracing of tree xylem water to test if the same water sources utilized by trees actively contributed to summer streamflow. We found that diel fluctuations in streamflow were not generated exclusively in the near-stream zone. We also found that the soil water sources trees utilized for transpiration were not directly linked to streamflow generation. This finding that the stored water utilized by the forest is different than the stored water contributing to streamflow has significant implications for how we conceptualize the interactions between forests and the hydrological system and how they are represented in hydrological models used to predict the response to forest management practices.

Our results provide new process-insights to how water is stored, extracted, and discharged from our forested catchments in Western Oregon that will help better explain how forest removal influences streamflow and water quality across multiple scales and geological conditions. Our approach of integrating an ongoing paired-catchment experiment with a process-based study represents a fruitful way forward in overcoming the challenge of predicting the hydrological effects of forest harvesting and other land management practices outside of the research setting.

## REFERENCES

Andréassian, V. (2004), Waters and forests: from historical controversy to scientific debate, *J. Hydrol.*, *291*(1-2), 1-27.

Bond, B. J., J. A. Jones, G. Moore, N. Phillips, D. Post, and J. J. McDonnell (2002), The zone of vegetation influence on baseflow revealed by diel patterns of streamflow and vegetation water use in a headwater basin, *Hydrol. Processes*, *16*(8), 1671-1677.

McGuire, K. J., J. J. McDonnell, M. Weiler, C. Kendall, B. L. McGlynn, J. M. Welker, and J. Seibert (2005), The role of topography on catchment-scale water residence time, *Water Resour. Res.*, *41*(5), -.

# References

**REFERENCES**


Ali, G. A., C. L'Heureux, A. G. Roy, M.-C. Turmel, and F. Courchesne (2011), Linking spatial patterns of perched groundwater storage and stormflow generation processes in a headwater forested catchment, *Hydrol. Processes*, n/a-n/a.

Amoozegar, A. (1989), A compact constant-head permeameter for measuring hydraulic conductivity of the vadose zone, *Soil Sci. Soc. Am. J.*, *53*, 1356-1361.

Anderson, S. P., W. E. Dietrich, R. Torres, D. R. Montgomery, and K. Loague (1997a), Concentration-discharge relationships in runoff from a steep, unchanneled catchment., *Water Resour. Res.*, *33*(1), 211-225.

Anderson, S. P., W. E. Dietrich, D. R. Montgomery, R. Torres, M. E. Conrad, and K. Loague (1997b), Subsurface flow paths in a steep, unchanneled catchment., *Water Resour. Res.*, *33*(12), 2637-2653.

Andreassian, V. (2004), Waters and forests: from historical controversy to scientific debate, *Journal of Hydrology*, *291*(1-2), 1-27.

Andréassian, V. (2004), Waters and forests: from historical controversy to scientific debate, *J. Hydrol.*, *291*(1-2), 1-27.

Arnold, J. G., P. M. Allen, R. Muttiah, and G. Bernhardt (1995), Automated Base Flow Separation and Recession Analysis Techniques, *Ground Water*, *33*(6), 1010-1018.

Asano, Y., T. Uchida, and N. Ohte (2002), Residence times and flow paths of water in steep unchannelled catchments, Tanakami, Japan, *J. Hydrol.*, *261*(1-4), 173-192.

Asbjornsen, H., et al. (2011), Ecohydrological advances and applications in plant–water relations research: a review, *Journal of Plant Ecology*, *4*(1-2), 3-22.

Baker, D. B., R. P. Richards, T. T. Loftus, and J. W. Kramer (2004), A NEW FLASHINESS INDEX: CHARACTERISTICS AND APPLICATIONS TO MIDWESTERN RIVERS AND STREAMS1, *JAWRA Journal of the American Water Resources Association*, *40*(2), 503-522.

Barnard, H. R., C. B. Graham, W. J. Van Verseveld, J. R. Brooks, B. J. Bond, and J. J. McDonnell (2010), Mechanistic assessment of hillslope transpiration controls of diel subsurface flow: a steady-state irrigation approach, *Ecohydrology*, *3*(2), 133-142.

Beschta, R. L. (1998), Forest hydrology in the Pacific Northwest: Additional research needs, *J. Am. Water Resour. Assoc.*, *34*(4), 729-742.

Beschta, R. L., M. R. Pyles, A. E. Skaugset, and C. G. Surfleet (2000), Peakflow responses to forest practices in the western cascades of Oregon, USA, *Journal of Hydrology*, *233*(1-4), 102-120.

Beven, K., and I. Westerberg (2011), On red herrings and real herrings: disinformation and information in hydrological inference, *Hydrol. Processes*, n/a-n/a.

Beven, K. J. (2001), How far can we go in distributed hydrological modelling?, *Hydrol. Earth Syst. Sci.*, *5*(1), 1-12.

Beven, K. J., and M. J. Kirkby (1979a), A physically-based variable contributing area model of basin hydrology, *Hydrological Sciences Bulletin*, *24*, 43-69.

Beven, K. J., and M. J. Kirkby (1979b), A physically based, variable contributing area model of basin hydrology., *Hydrological Sciences Bulletin*, *24*(1), 43-69.

Bierlmaler, F. A., and A. McKee (1989), Climatic summaries and documentation for the primary meteorological station, H.J. Andrews Experimental Forest, 1972 to 1984, *U.S. Department of Agriculture, Forest Service, Pacific Northwest Research Station*, *PNW-GTR-242*, 61.

Birkel, C., C. Soulsby, and D. Tetzlaff (2011a), Modelling catchment-scale water storage dynamics: reconciling dynamic storage with tracer-inferred passive storage, *Hydrol. Processes*, n/a-n/a.

Birkel, C., D. Tetzlaff, S. M. Dunn, and C. Soulsby (2011b), Using time domain and geographic source tracers to conceptualize streamflow generation processes in lumped rainfall-runoff models, *Water Resour. Res.*, *47*(2), W02515.

Blöschl, G. (2001), Scaling in hydrology, *Hydrol. Processes*, *15*, 709-711.

Bond, B. J., J. A. Jones, G. Moore, N. Phillips, D. Post, and J. J. McDonnell (2002), The zone of vegetation influence on baseflow revealed by diel patterns of streamflow and vegetation water use in a headwater basin, *Hydrol. Processes*, *16*(8), 1671-1677.

Bonell, M. (1993), Progress in the understanding of runoff generation dynamics in forests, *J. Hydrol.*, *150*, 217-275.

Bonell, M., and L. A. Bruijnzeel (Eds.) (2004), *Forests, Water and People in the Humid Tropics: Past, Present and Future Hydrological Research for Integrated Land and Water Management*, Cambridge University Press, Cambridge.

Bosch, J. M., and J. D. Hewlett (1982), A REVIEW OF CATCHMENT EXPERIMENTS TO DETERMINE THE EFFECT OF VEGETATION CHANGES ON WATER YIELD AND EVAPO-TRANSPIRATION, *Journal of Hydrology*, *55*(1-4), 3-23.

Botter, G., E. Bertuzzo, and A. Rinaldo (2010), Transport in the hydrologic response: Travel time distributions, soil moisture dynamics, and the old water paradox, *Water Resour. Res.*, *46*.

Botter, G., A. Porporato, I. Rodriguez-Iturbe, and A. Rinaldo (2009), Nonlinear storage-discharge relations and catchment streamflow regimes, *Water Resour. Res.*, *45*.

Brooks, J. R., H. R. Barnard, R. Coulombe, and J. J. McDonnell (2010), Ecohydrologic separation of water between trees and streams in a Mediterranean climate, *Nature Geoscience*, *3*(2), 100-104.

Brown, G. W. (1969), Predicting temperatures of small streams, *Water Resour. Res.*, *5*(1), 68-75.

Broxton, P. D., P. A. Troch, and S. W. Lyon (2009), On the role of aspect to quantify water transit times in small mountainous catchments, *Water Resour. Res.*, *45*.

Brutsaert, W., and J. L. Nieber (1977), REGIONALIZED DROUGHT FLOW HYDROGRAPHS FROM A MATURE GLACIATED PLATEAU, *Water Resour. Res.*, *13*(3), 637-644.

Butler, J. J., Jr., G. J. Kluitenberg, D. O. Whittemore, S. P. Loheide, II, W. Jin, M. A. Billinger, and X. Zhan (2007), A field investigation of phreatophyte-induced fluctuations in the water table, *Water Resour. Res.*, *43*(2), W02404.

Calder, I. R. (1998), Water use by forests, limits and controls, *Tree Physiology*, *18*(8-9), 625-631.

Capell, R., D. Tetzlaff, A. J. Hartley, and C. Soulsby (2011a), Linking metrics of hydrological function and transit times to landscape controls in a heterogeneous mesoscale catchment, *Hydrol. Processes*, n/a-n/a.

Capell, R., D. Tetzlaff, I. A. Malcolm, A. J. Hartley, and C. Soulsby (2011b), Using hydrochemical tracers to conceptualise hydrological function in a larger scale catchment draining contrasting geologic provinces, *J. Hydrol.*, *408*(1-2), 164-177.

Constantz, J. (1998), Interaction between stream temperature, streamflow, and groundwater exchanges in alpine streams., *Water Resour. Res.*, *34*(7), 1609-1615.

Corliss, J. F. (1973), Soil Survey, Alsea Area, Oregon, *U.S. Soil Conservation Service*, 82 pp.

Cuevas, J. G., M. Calvo, C. Little, M. Pino, and P. Dassori (2010), ARE DIURNAL FLUCTUATIONS IN STREAM FLOW REAL?, *Journal of Hydrology and Hydromechanics*, *58*(3), 149-162.

Dawson, T. E., and J. R. Ehleringer (1991), Streamside trees that do not use stream water, *Nature*, *350*(6316), 335-337.

Descroix, L., J. F. Nouvelot, and M. Vauclin (2002), Evaluation of an antecedent precipitation index to model runoff yield in the western Sierra Madre (North-west Mexico), *J. Hydrol.*, *263*(1-4), 114-130.

DeWalle, D. R., P. J. Edwards, B. R. Swistock, R. Aravena, and R. J. Drimmie (1997), Seasonal isotope hydrology of three Appalachian forest catchments, *Hydrol. Processes*, *11*(15), 1895-1906.

Dietrich, W. E., and T. Dunne (1978), Sediment budget for a small catchment in mountainous terrain, *Z. Geomorph. Suppl.*, *29*, 191-206.

Dingman, S. L. (2002), *Physical hydrology*, 2nd ed., x, 646 p. pp., Prentice Hall, Upper Saddle River, N.J.

Dooge, J. C. I. (1986), Looking for Hydrologic Laws, *Water Resour. Res.*, *22*, 46S-58S.

Dunford, E. G., and P. W. Fletcher (1947), Effect of removal of stream-bank vegetation upon water yield, *Eos Trans. AGU*, *28*, 105-110.

Ehleringer, J. R., and C. B. Osmond (1989), Stable Isotopes, in *Plant Physiological Ecology Field Methods and Instrumentation*, edited by R. W. Pearcy, J. R. Ehleringer, H. A. Mooney and P. W. Rundel, pp. 281-300, Chapman and Hall, Ltd., London.

Ehleringer, J. R., and T. E. Dawson (1992), WATER-UPTAKE BY PLANTS - PERSPECTIVES FROM STABLE ISOTOPE COMPOSITION, *Plant Cell and Environment*, *15*(9), 1073-1082.

Ellsworth, P. Z., and D. G. Williams (2007), Hydrogen isotope fractionation during water uptake by woody xerophytes, *Plant and Soil*, *291*(1-2), 93-107.

Fenicia, F., S. Wrede, D. Kavetski, L. Pfister, L. Hoffmann, H. H. G. Savenije, and J. J. McDonnell (2010), Assessing the impact of mixing assumptions on the estimation of streamwater mean residence time, *Hydrol. Processes*, *24*(12), 1730-1741.

Freeze, R. A., and J. A. Cherry (1979), *Groundwater*, 1st ed., 604 pp., Prentice-Hall, Englewood Cliffs, NJ.

Gabrielli, C. P., and J. J. McDonnell (2011), An inexpensive and portable drill rig for bedrock groundwater studies in headwater catchments, *Hydrol. Processes*, n/a-n/a.

Gardner, P., and D. K. Solomon (2009), An advanced passive diffusion sampler for the determination of dissolved gas concentrations, *Water Resour. Res.*, *45*.

Gat, J., and Y. Tzur (1967), Modification of the isotopic composition of rainwater by processes which occur before groundwater recharge, in *Proceedings of the Symposium on Isotopes in Hydrology*, edited, pp. 49-60, International Atomic Energy Agency, Vienna, Austria, 14-18 November 1966.

Gippel, C. J., and M. J. Stewardson (1998), Use of wetted perimeter in defining minimum environmental flows., *Regulated Rivers: Research and Management*, *14*, 53-67.

Gleeson, T., L. Smith, N. Moosdorf, J. Hartmann, H. H. Dürr, A. H. Manning, L. P. H. van Beek, and A. M. Jellinek (2011), Mapping permeability over the surface of the Earth, *Geophys. Res. Lett.*, *38*(2), L02401.

Goldsmith, G. R., L. E. Muñoz-Villers, F. Holwerda, J. J. McDonnell, H. Asbjornsen, and T. E. Dawson (2011), Stable isotopes reveal linkages among ecohydrological processes in a seasonally dry tropical montane cloud forest, *Ecohydrology*, n/a-n/a.

Gomi, T., R. C. Sidle, and J. S. Richardson (2002), Understanding processes and downstream linkages of headwater systems, *Bioscience*, *52*(10), 905–916.

Graham, C. B., and J. J. McDonnell (2010), Hillslope threshold response to rainfall: (2) Development and use of a macroscale model, *J. Hydrol.*, *393*(1-2), 77-93.

Granier, A. (1987), Evaluation of transpiration in a Douglas-fir stand by means of sap flow measurements, *Tree Physiology*, *3*, 309-320.

Gribovszki, Z., J. Szilagyi, and P. Kalicz (2010), Diurnal fluctuations in shallow groundwater levels and streamflow rates and their interpretation - A review, *J. Hydrol.*, *385*(1-4), 371-383.

Gribovszki, Z., P. Kalicz, J. Szilagyi, and M. Kucsara (2008), Riparian zone evapotranspiration estimation from diurnal groundwater level fluctuations, *J. Hydrol.*, *349*(1-2), 6-17.

Groom, J. D., L. Dent, and L. J. Madsen (2011), Stream temperature change detection for state and private forests in the Oregon Coast Range, *Water Resour. Res.*, *47*(1), W01501.

Hale, V. C., and J. J. McDonnell (this issue), Similar catchment forms and rainfall-runoff responses hide radically different plumbing: (1) A top-down analysis of mean transit time ranges and scaling relations, *to be submitted to Water Resources Research*.

Hale, V. C., J. J. McDonnell, M. K. Stewart, D. K. Solomon, G. G. Ice, and R. T. Pack (this issue), Similar catchment form and rainfall-runoff responses hide radically different plumbing: (2) The role of catchment storage controls, *to be submitted to Water Resources Research*.

Haria, A. H., and P. Shand (2004), Evidence for deep subsurface flow routing in a forested upland Wales: implications for contaminant transport and stream flow generation, *Hydrol. Earth Syst. Sci.*, *8*(2), 334-344.

Harr, R. D. (1976a), Forest practices and streamflow in western Oregon. Gen. Tech. Rep. PNW-49*Rep.*, 18 pp, U.S. Department of Agriculture, Forest Service, Pacific Northwest Forest and Range Experiment Station, Portland, OR.

Harr, R. D. (1976b), Hydrology of small forest stream in western Oregon, General Technical Report PNW-55*Rep.*, 15 pp, Pacific Northwest Forest and Range Experiment Station, USDA Forest Service, Portland, OR.

Harr, R. D. (1977), Water flux in soil and subsoil on a steep forested slope, *J. Hydrol.*, *33*, 37-58.

Harr, R. D. (1981), Some characteristics and consequences of snowmelt during rainfall in western, Oregon, *J. Hydrol.*, *53*, 277-304.

Harrington, C. A., J. C. Zasada, and E. A. Allen (1994), Biology of red alder (*Alnus rubra* Bong.), in *The Biology and Management of Red Alder*, edited by D. E. Hibbs, D. S. DeBell and R. F. Tarrant, pp. 3-22, Oregon State University Press, Corvallis, OR

Harris, D. D. (1977), Hydrologic changes after logging in two small Oregon coastal watersheds, *United States Geological Survey*, *Water-Supply Paper 2037*.

Hewlett, J. D. (1982), *Principles of Forest Hydrology*, 183 pp., University of Georgia Press, Athens, GA.

Hewlett, J. D., and A. R. Hibbert (1967), Factors affecting the response of small watersheds to precipitation in humid areas, in *Forest Hydrology.*, edited by W. E. Sopper and H. W. Lull, pp. 275-291, Pergamon Press, New York.

Hewlett, J. D., H. W. Lull, and K. G. Reinhart (1969), In Defense of Experimental Watersheds, *5*(1), 306-316.

Hibbert, A. R. (1967), Forest treatment effects in water yield, paper presented at Forest Hydrology, Pergamon Press, The Pennsylvania State University, University Park, PA, Aug. 29 to Sept. 10, 1965.

Hjerdt, K. N., J. J. McDonnell, J. Seibert, and A. Rodhe (2004), A new topographic index to quantify downslope controls on local drainage, *Water Resour. Res.*, *40*, W05602, doi:05610.01029/02004WR003130.

Hrachowitz, M., C. Soulsby, D. Tetzlaff, J. J. C. Dawson, and I. A. Malcolm (2009), Regionalization of transit time estimates in montane catchments by integrating landscape controls, *Water Resour. Res.*, *45*.

Hrachowitz, M., C. Soulsby, D. Tetzlaff, I. A. Malcolm, and G. Schoups (2010), Gamma distribution models for transit time estimation in catchments: Physical interpretation of parameters and implications for time-variant transit time assessment, *Water Resour. Res.*, *46*.

Ice, G. G., V. C. Hale, J. Light, J. J. McDonnell, S. H. Schoenholtz, and J. D. Stednick (2007), Testing contemporary forest practices: The Alsea Watershed Study Revisited., in *Sustaining America's Forests: Proceedings of the 2007 Society of American Foresters National Convention*, edited, Society of American Foresters, Portland, OR.

Ingebritsen, S. E., D. R. Sherrod, and R. H. Mariner (1992), RATES AND PATTERNS OF GROUNDWATER-FLOW IN THE CASCADE RANGE VOLCANIC ARC, AND THE EFFECT ON SUBSURFACE TEMPERATURES, *Journal of Geophysical Research-Solid Earth*, *97*(B4), 4599-4627.

Jefferson, A., A. Nolin, S. Lewis, and C. Tague (2008), Hydrogeologic controls on streamflow sensitivity to climate variation, *Hydrol. Processes*, *22*(22), 4371-4385.

Jefferson, A., G. E. Grant, S. L. Lewis, and S. T. Lancaster (2010), Coevolution of hydrology and topography on a basalt landscape in the Oregon Cascade Range, USA, *Earth Surf. Processes Landforms*, *35*(7), 803-816.

Jenny, H. (1941), *Factors of Soil Formation*, Dover Publications, Inc., New York.

Jones, J. A., and G. E. Grant (1996), Peak flow responses to clear-cutting and roads in small and large basins, western Cascades, Oregon, *Water Resour. Res.*, *32*(4), 959-974.

Jones, J. A., and D. A. Post (2004), Seasonal and successional streamflow response to forest cutting and regrowth in the northwest and eastern United States, *Water Resour. Res.*, *40*(W05203), doi:10.1029/2003WR002952.

Kabeya, N., M. Katsuyama, M. Kawasaki, N. Ohte, and A. Sugimoto (2007), Estimation of mean residence times of subsurface waters using seasonal variation in deuterium excess in a small headwater catchment in Japan, *Hydrol. Processes*, *21*, 308-322.

Katsuyama, M., M. Tani, and S. Nishimoto (2010), Connection between streamwater mean residence time and bedrock groundwater recharge/discharge dynamics in weathered granite catchments, *Hydrol. Processes*, *24*(16), 2287-2299.

Kibler, K. M. (2007), The influence of contemporary forest harvesting on summer stream temperatures in headwater streams of Hinkle Creek, Oregon, 115 pp, MS Thesis, Oregon State University, Corvallis, OR.

Kirchner, J. W. (2006), Getting the right answers for the right reasons: Linking measurements, analyses, and models to advance the science of hydrology, *Water Resour. Res.*, *42*.

Kirchner, J. W. (2009), Catchments as simple dynamical systems: Catchment characterization, rainfall-runoff modeling, and doing hydrology backward, *Water Resour. Res.*, *45*.

Kirchner, J. W., X. Feng, and C. Neal (2000), Fractal stream chemistry and its implications for contaminant transport in catchments, *Nature*, *403*(6769), 524-527.

Kosugi, K. i., S. y. Katsura, M. Katsuyama, and M. Takahisa (2006), Water flow processes in weathered granitic bedrock and their effects on runoff generation in a small headwater catchment, *Water Resour. Res.*, *42*.

Kosugi, K. i., M. Fujimoto, S. y. Katsura, H. Kato, Y. Sando, and T. Mizuyama (2011), Localized bedrock aquifer distribution explains discharge from a headwater catchment, *Water Resour. Res.*, *47*(7), W07530.

Kovner, J. L., and T. C. Evans (1954), A Method for Determining the Minimum Duration of Watershed Experiments, *Transactions, American Geophysical Union*, *35*(4), 608-612.

Lin, G. H., and L. D. L. Sternberg (1993), EFFECTS OF SALINITY FLUCTUATION ON PHOTOSYNTHETIC GAS-EXCHANGE AND PLANT-GROWTH OF THE RED MANGROVE (RHIZOPHORA-MANGLE L), *J. Exp. Bot.*, *44*(258), 9-16.

Lohse, K. A., and W. E. Dietrich (2005), Contrasting effects of soil development on hydrological properties and flow paths, *Water Resour. Res.*, *41*(12).

Lundquist, J. D., and D. R. Cayan (2002), Seasonal and spatial patterns in diurnal cycles in streamflow in the western United States, *Journal of Hydrometeorology*, *3*(5), 591-603.

Maloszewski, P., and A. Zuber (1982), Determining the turnover time of groundwater systems with the aid of environmental tracers. 1. models and their applicability, *J. Hydrol.*, *57*, 207-231.

Manning, A. H., and J. S. Caine (2007), Groundwater noble gas, age, and temperature signatures in an Alpine watershed: Valuable tools in conceptual model development, *Water Resour. Res.*, *43*(4).

Marcus, H., and D. E. Evenson (1961), Directional permeability in anisotropic porous media, *Water Resources Center Contrib. No. 31*, University of California, Berkeley.

Mazurkiewicz, A. B., D. G. Callery, and J. J. McDonnell (2008), Assessing the controls of the snow energy balance and water available for runoff in a rain-on-snow environment, *J. Hydrol.*, *354*(1-4), 1-14.

McDonnell, J. J., and T. Tanaka (2001), Hydrology and biogeochemistry of forested catchments., in *Hydrol. Processes*, edited, John Wiley & Sons, New York.

McDonnell, J. J., and R. Woods (2004), On the need for catchment classification, *J. Hydrol.*, *299*(1-2), 2-3.

McDonnell, J. J., et al. (2010), How old is streamwater? Open questions in catchment transit time conceptualization, modelling and analysis, *Hydrol. Processes*, *24*(12), 1745-1754.

McGlynn, B., J. McDonnell, M. Stewart, and J. Seibert (2003), On the relationships between catchment scale and streamwater mean residence time, *Hydrol. Processes*, *17*(1), 175-181.

McGlynn, B. L., and J. J. McDonnell (2003), Quantifying the relative contributions of riparian and hillslope zones to catchment runoff, *Water Resour. Res.*, *39*(11), 1310, doi:1310.1029/2003WR002091.

McGlynn, B. L., and J. Seibert (2003), Distributed assessment of contributing area and riparian buffering along stream networks, *Water Resour. Res.*, *39*(4), 1082, doi:1010.1029/2002WR001521.

McGuire, K. J., and J. J. McDonnell (2006), A review and evaluation of catchment transit time modeling, *J. Hydrol.*, *330*(3-4), 543-563.

McGuire, K. J., and J. J. McDonnell (2010), Hydrological connectivity of hillslopes and streams: Characteristic time scales and nonlinearities, *Water Resour. Res.*, *46*.

McGuire, K. J., J. J. McDonnell, M. Weiler, C. Kendall, B. L. McGlynn, J. M. Welker, and J. Seibert (2005), The role of topography on catchment-scale water residence time, *Water Resour. Res.*, *41*(5), -.

McMinn, R. G. (1963), CHARACTERISTICS OF DOUGLAS-FIR ROOT SYSTEMS, *Canadian Journal of Botany*, *41*(1), 105-122.

McNamara, J. P., D. Tetzlaff, K. Bishop, C. Soulsby, M. Seyfried, N. E. Peters, B. T. Aulenbach, and R. Hooper (2011), Storage as a metric of catchment comparison, *Hydrol. Processes*, n/a-n/a.

Meinzer, O. E. (1927), Plants as indicators of ground water, *U.S. Geological Survey Water-Supply Paper 577*, 95.

Millares, A., M. J. Polo, and M. A. Losada (2009), The hydrological response of baseflow in fractured mountain areas, *Hydrol. Earth Syst. Sci.*, *13*(7), 1261-1271.

Molina, R., D. Myrold, and C. Y. Li (1994), Root symbioses of red alder: Technological opportunities for enhanced regeneration and soil improvement, in *The Biology and Management of Red Alder*, edited by D. E. Hibbs, D. S. DeBell and R. F. Tarrant, pp. 23-46, Oregon State University Press, Corvallis, OR

Monteith, D. T. (1965), Evaporation and the environment, *Symposium of the Society for Experimental Biology*, *19*, 205-234.

Montgomery, D. R., W. E. Dietrich, and J. T. Heffner (2002), Piezometric response in shallow bedrock at CB1: Implications for runoff generation and landsliding, *Water Resour. Res.*, *38*(12), doi:10.1029/2002WR001429.

Montgomery, D. R., W. E. Dietrich, R. Torres, S. P. Anderson, J. T. Heffner, and K. Loague (1997), Hydrologic response of a steep, unchanneled valley to natural and applied rainfall, *Water Resour. Res.*, *33*(1), 91-109.

Moore, G. W., B. J. Bond, J. A. Jones, N. Phillips, and F. C. Meinzer (2004), Structural and compositional controls on transpiration in 40-and 450-year-old riparian forests in western Oregon, USA, *Tree Physiology*, *24*(5), 481-491.

Morgenstern, U. (2007), Lake Taupo Streams: Water age distribution, fraction of landuse impacted water, and future nitrogen load*Rep.*, 21 pp, Environment Waikato, Hamilton, New Zealand.

Morgenstern, U., and C. B. Taylor (2009), Ultra low-level tritium measurement using electrolytic enrichment and LSC, *Isotopes in Environmental and Health Studies*, *45*(2), 96-117.

Morgenstern, U., R. R. Reeves, C. J. Daughney, S. Cameron, and D. Gordon (2005), Groundwater age and chemistry, and future nutrient load for selected Rotorua lakes catchments*Rep.*, 73 pp.

Murtaugh, P. A. (2000), Paired intervention analysis in ecology, *Journal of Agricultural Biological and Environmental Statistics*, *5*(3), 280-292.

Niemi, A. J. (1978), Residence time distribution of variable flow processes, *International Journal of Applied Radiation and Isotopes*, *28*, 855-860.

NRC (2008), *Hydrologic Effects of a Changing Forest Landscape*, National Acadamies Press, Washington, D.C.

Olden, J. D., and N. L. Poff (2003), Redundancy and the choice of hydrologic indices for characterizing streamflow regimes, *River Res. Appl.*, *19*(2), 101-121.

Onda, Y., M. Tsujimura, J.-i. Fujihara, and J. Ito (2006), Runoff generation mechanisms in high-relief mountainous watersheds with different underlying geology, *J. Hydrol.*, *331*(3-4), 659-673.

Oswald, C. J., M. C. Richardson, and B. A. Branfireun (2011), Water storage dynamics and runoff response of a boreal Shield headwater catchment, *Hydrol. Processes*, *25*(19), 3042-3060.

Oudin, L., A. Kay, V. Andréassian, and C. Perrin (2010), Are seemingly physically similar catchments truly hydrologically similar?, *Water Resour. Res.*, *46*(11), W11558.

Peck, D. V., A. B. Griggs, H. G. Schlicker, F. G. Wells, and H. M. Dole (1964), Geology of the central and nothern parts of the western Cascade Range in Oregon, *U.S. Geological Survey Professional Paper 449*, 56.

Penman, H. L. (1948), Natural evaporation from open water, bare soil, and grass, *Proceedings of the Royal Society of London*, *A193*, 120-146.

Raats, P. A. C. (2007), Uptake of water from soils by plant roots, *Transport in Porous Media*, *68*(1), 5-28.

Ranken, D. W. (1974), Hydrologic properties of soil and subsoil on a steep, forested slope, M.S. thesis, 114 pp, Oregon State University, Corvallis.

Rasmussen, T. C. (2001), Pressure wave vs. tracer velocities through unsaturated fractured rock, *Flow and transport through unsaturated fractured rock; an overview*, *42*, 45-52.

Reager, J. T., and J. S. Famiglietti (2009), Global terrestrial water storage capacity and flood potential using GRACE, *Geophys. Res. Lett.*, *36*.

Rinaldo, A., K. J. Beven, E. Bertuzzo, L. Nicotina, J. Davies, A. Fiori, D. Russo, and G. Botter (2011), Catchment travel time distributions and water flow in soils, *Water Resour. Res.*, *47*.

Robinson, D. A., et al. (2008), Advancing process-based watershed hydrological research using near-surface geophysics: a vision for, and review of, electrical and magnetic geophysical methods, *Hydrol. Processes*, *22*(18), 3604-3635.

Rodgers, P., C. Soulsby, and D. Tetzlaff (2005), Using stable isotope tracers to assess hydrological flow paths, residence times and landscape influences in a nested mesoscale catchment, *Hydrol. Earth Syst. Sci.*, *9*, 139-155.

Roering, J. J., J. W. Kirchner, and W. E. Dietrich (2005), Characterizing structural and lithologic controls on deep-seated landsliding: Implications for topographic relief and landscape evolution in the Oregon Coast Range, USA, *Geological Society of America Bulletin*, *117*(5-6), 654-668.

Rowe, P. B. (1963), Streamflow increases after removing woodland-riparian vegetation from a southern California watershed, *Journal of Forestry*, *365-370*.

Rupp, D. E., and R. A. Woods (2008), Increased flexibility in base flow modelling using a power law transmissivity profile, *Hydrol. Processes*, *22*(14), 2667-2671.

Saar, M. O., and M. Manga (2004), Depth dependence of permeability in the Oregon Cascades inferred from hydrogeologic, thermal, seismic, and magmatic modeling constraints, *J. Geophys. Res.*, *109*(B4), B04204.

Sawicz, K., T. Wagener, M. Sivapalan, P. A. Troch, and G. Carrillo (2011), Catchment classification: empirical analysis of hydrologic similarity based on catchment function in the eastern USA, *Hydrol. Earth Syst. Sci.*, *15*(9), 2895-2911.

Sayama, T., and J. J. McDonnell (2009), A new time-space accounting scheme to predict stream water residence time and hydrograph source components at the watershed scale, *Water Resour. Res.*, *45*.

Sayama, T., J. J. McDonnell, A. Dhakal, and K. Sullivan (2011), How much water can a watershed store?, *Hydrol. Processes*, n/a-n/a.

Schlosser, P., M. Stute, H. Dorr, C. Sonntag, and K. O. Munnich (1988), TRITIUM HE-3 DATING OF SHALLOW GROUNDWATER, *Earth and Planetary Science Letters*, *89*(3-4), 353-362.

Schoups, G., and J. A. Vrugt (2010), A formal likelihood function for parameter and predictive inference of hydrologic models with correlated, heteroscedastic, and non-Gaussian errors, *Water Resour. Res.*, *46*.

Seibert, J., and J. J. McDonnell (2010), Land-cover impacts on streamflow: a change-detection modelling approach that incorporates parameter uncertainty, *Hydrol. Sci. J.-J. Sci. Hydrol.*, *55*(3), 316-332.

Sidle, R. C. (2006), Field observations and process understanding in hydrology: essential components in scaling, *Hydrol. Processes*, *20*, 1439-1445.

Sidle, R. C., S. Noguchi, Y. Tsuboyama, and K. Laursen (2001), A conceptual model of preferential flow systems in forested hillslopes: evidence of self-organization, *Hydrol. Processes*, *15*, 1675-1692.

Singh, V. P., and D. A. Woolhiser (2002), Mathematical modeling of watershed hydrology, *Journal of Hydrologic Engineering*, *7*(4), 270-292.

Sivapalan, M. (2003), Process complexity at hillslope scale, process simplicity at the watershed scale: is there a connection?, *Hydrol. Processes*, *17*(5), 1037-1041.

Sivapalan, M. (2005), Pattern, process, and function: Elements of a unified theory of hydrology at the catchment scale, in *Encyclopedia of Hydrological Sciences*, edited by M. G. Anderson, pp. 193-219, John Wiley & Sons, Ltd.

Sivapalan, M. (2009), The secret to 'doing better hydrological science': change the question!, *Hydrol. Processes*, *23*(9), 1391-1396.

Smakhtin, V. U. (2001), Low flow hydrology: a review, *J. Hydrol.*, *240*, 3-4.

Snavely, P. D., H. C. Wagner, and N. S. MacLeod (1964), Rythmic-bedded eugeosynclinal deposits of the Tyee Formation, Oregon Coast Range, in *Symposium on cyclic sedimentation: Kansas Geological Survey, Bulletin 169, pp. 461-480*, edited by D. F. Merriam.

Solomon, D. K., and P. G. Cook (2000), $^3$H and $^3$He, in *Environmental Tracers in Subsurface Hydrology*, edited by P. Cook and A. L. Herczeg, pp. 397-424, Kluwer Academic Publishers, Boston.

Solomon, D. K., D. P. Genereux, L. N. Plummer, and E. Busenberg (2010), Testing mixing models of old and young groundwater in a tropical lowland rain forest with environmental tracers, *Water Resour. Res.*, *46*.

Soulsby, C., D. Tetzlaff, and M. Hrachowitz (2009), Tracers and transit times: windows for viewing catchment scale storage?, *Hydrol. Processes*, *23*(24), 3503-3507.

Soulsby, C., D. Tetzlaff, and M. Hrachowitz (2010), Are transit times useful process-based tools for flow prediction and classification in ungauged basins in montane regions?, *Hydrol. Processes*, *24*(12), 1685-1696.

Soulsby, C., D. Tetzlaff, S. M. Dunn, and S. Waldron (2006a), Scaling up and out in runoff process understanding: insights from nested experimental catchment studies, *Hydrol. Processes*, *20*(11), 2461-2465.

Soulsby, C., R. Malcolm, R. C. Ferrier, R. C. Helliwell, and A. Jenkins (2000), Isotope hydrology of the Allt a'Mharcaidh catchment, Cairngorms, Scotland: implications for hydrological pathways and residence times, *Hydrol. Processes*, *14*, 747-762.

Soulsby, C., D. Tetzlaff, P. Rodgers, S. Dunn, and S. Waldron (2006b), Runoff processes, stream water residence times and controlling landscape characteristics in a mesoscale catchment: An initial evaluation, *J. Hydrol.*, *325*(1-4), 197-221.

Soulsby, C., D. Tetzlaff, N. van den Bedem, I. A. Malcolm, P. J. Bacon, and A. F. Youngson (2007), Inferring groundwater influences on surface water in montane catchments from hydrochemical surveys of springs and streamwaters, *J. Hydrol.*, *333*(2-4), 199-213.

Spence, C. (2010), A Paradigm Shift in Hydrology: Storage Thresholds Across Scales Influence Catchment Runoff Generation, *Geography Compass*, *4*(7), 819-833.

Stanfield, B. J., J. C. Bliss, and T. A. Spies (2002), Land ownership and landscape structure: a spatial analysis of sixty-six Oregon (USA) Coast Range watersheds, *Landscape Ecology*, *17*(8), 685-697.

Stednick, J. D. (1996), Monitoring the effects of timber harvest on annual water yield, *Journal of Hydrology (Amsterdam)*, *176*(1/4), 79-95.

Stednick, J. D. (Ed.) (2008), *Hydrological and Biological Responses to Forest Practices: The Alsea Watershed Study*, Springer, New York.

Sternberg, L. d. S. L., and P. K. Swart (1987), Utilization of Freshwater and Ocean Water by Coastal Plants of Southern Florida, *Ecology*, *68*(6), 1898-1905.

Steudle, E. (2000), Water uptake by plant roots: an integration of views, *Plant and Soil*, *226*(1), 45-56.

Stewart, M. K., and J. J. McDonnell (1991), Modeling base flow soil water residence times from deuterium concentrations, *Water Resour. Res.*, *27*(10), 2681-2693.

Stewart, M. K., U. Morgenstern, and J. J. McDonnell (2010), Truncation of stream residence time: how the use of stable isotopes has skewed our concept of streamwater age and origin, *Hydrol. Processes*, *24*(12), 1646-1659.

Swanson, F. J., and M. E. James (1975), Geology and geomorphology of the H.J. Andrews Experimental Forest, western Cascades, Oregon.*Rep. Res. Pap. PNW-188*, 14 pp, U.S. Department of Agriculture, Forest Service, Pacific Northwest Forest and Range Experiment Station, Portland, OR.

Szilagyi, J., Z. Gribovszki, P. Kalicz, and M. Kucsara (2008), On diurnal riparian zone groundwater-level and streamflow fluctuations, *J. Hydrol.*, *349*(1-2), 1-5.

Tague, C., and G. E. Grant (2004), A geological framework for interpreting the low-flow regimes of Cascade streams, Willamette River Basin, Oregon, *Water Resour. Res.*, *40*(W04303), doi:10.1029/2003WR002629.

Tarboton, D. G. (1997), A new method for the determination of flow directions and upslope areas in grid digital elevation models, *Water Resour. Res.*, *33*, 309-319.

Tetzlaff, D., J. Seibert, and C. Soulsby (2009a), Inter-catchment comparison to assess the influence of topography and soils on catchment transit times in a geomorphic province; the Cairngorm mountains, Scotland, *Hydrol. Processes*, *23*(13), 1874-1886.

Tetzlaff, D., J. P. McNamara, and S. K. Carey (2011), Measurements and modelling of storage dynamics across scales, *Hydrol. Processes*, *25*(25), 3831-3835.

Tetzlaff, D., S. K. Carey, H. Laudon, and K. McGuire (2010), Catchment processes and heterogeneity at multiple scales—benchmarking observations, conceptualization and prediction, *Hydrol. Processes*, *24*(16), 2203-2208.

Tetzlaff, D., J. Seibert, K. J. McGuire, H. Laudon, D. A. Burns, S. M. Dunn, and C. Soulsby (2009b), How does landscape structure influence catchment transit time across different geomorphic provinces?, *Hydrol. Processes*, *23*(6), 945-953.

Tetzlaff, D., J. J. McDonnell, S. Uhlenbrook, K. J. McGuire, P. W. Bogaart, F. Naef, A. J. Baird, S. M. Dunn, and C. Soulsby (2008), Conceptualizing catchment processes: simply too complex?, *Hydrol. Processes*, *22*(11), 1727-1730.

Thomas, R. B., and W. F. Megahan (1998), Peak flow responses to clear-cutting and roads in small and large basins, western Cascades, Oregon: a second opinion, *Water Resources Research*, *34*(12), 3393-3403.

Thorson, T. D., S.A. Bryce, D.A. Lammers, A.J. Woods, J.M. Omernik, J. Kagan, D.E. Pater, and J.A. Comstock (2003), Ecoregions of Oregon, *U.S. Geological Survey, Reston, VA.*

Torres, R., W. E. Dietrich, D. R. Montgomery, S. P. Anderson, and K. Loague (1998), Unsaturated zone processes and the hydrologic response of a steep, unchanneled catchment, *Water Resour. Res.*, *34*(8), 1865-1879.

Troxell, M. C. (1936), The diurnal fluctuation in the groundwater and flow of the Santa Ana River and its meaning, *Trans. AGU*, *17*, 496-504.

Uchida, T., J. J. McDonnell, and Y. Asano (2006), Functional intercomparison of hillslopes and small catchments by examining water source, flowpath and mean residence time, *J. Hydrol.*, *327*, 627-642.

Vanclay, J. K. (2009), Managing water use from forest plantations, *For. Ecol. Manage.*, *257*(2), 385-389.

Wagener, T., M. Sivapalan, P. Troch, and R. Woods (2007), Catchment classification and hydrologic similarity, *Geography Compass*, *1*.

West, A. G., G. R. Goldsmith, P. D. Brooks, and T. E. Dawson (2010), Discrepancies between isotope ratio infrared spectroscopy and isotope ratio mass spectrometry for the stable isotope analysis of plant and soil waters, *Rapid Communications in Mass Spectrometry*, *24*(14), 1948-1954.

Western, A. W., R. B. Grayson, and G. Bloschl (2002), Scaling of soil moisture: A hydrologic perspective, *Annual Review of Earth and Planetary Sciences*, *30*, 149-180.

Weyman, D. R. (1973), Measurements of the downslope flow of water in a soil, *J. Hydrol.*, *20*, 267-288.

Whipkey, R. Z. (1965), Subsurface storm flow from forested slopes, *Bull. Int. Assoc. Sci. Hydrol.*, *2*, 74-85.

White, W. N. (1932), A method of estimating ground-water supplies based on discharge by plants and evaporation from soil: Results of investigations in Escalante Valley, Utah, *U.S. Geological Survey Water-supply Paper 659-A*, 105.

Wicht, C. L. (1941), Diurnal fluctuations in Jonkershoek streams due to evaporation and transpiration, *Journal of the South African Forestry Association*, *7*, 34-49.

Wilson, C. J., and W. E. Dietrich (1987), The contribution of bedrock groundwater flow to storm runoff and high pore pressure development in hollows, in *Erosion and Sedimentation in the Pacific Rim*, edited, IAHS Publication No. 165, Corvallis, OR.

Wondzell, S. M. (2006), Effect of morphology and discharge on hyporheic exchange flows in two small streams in the Cascade Mountains of Oregon, USA, *Hydrol. Processes*, *20*(2), 267-287.

Wondzell, S. M., M. N. Gooseff, and B. L. McGlynn (2007), Flow velocity and the hydrologic behavior of streams during baseflow, *Geophys. Res. Lett.*, *34*(24).

Wondzell, S. M., M. N. Gooseff, and B. L. McGlynn (2010), An analysis of alternative conceptual models relating hyporheic exchange flow to diel fluctuations in discharge during baseflow recession, *Hydrol. Processes*, *24*(6), 686-694.

Yadav, M., T. Wagener, and H. Gupta (2007), Regionalization of constraints on expected watershed response behavior for improved predictions in ungauged basins, *Adv. Water Resour.*, *30*(8), 1756-1774.

Zegre, N., A. E. Skaugset, N. A. Som, J. J. McDonnell, and L. M. Ganio (2010), In lieu of the paired catchment approach: Hydrologic model change detection at the catchment scale, *Water Resour. Res.*, *46*.