University of Alberta

# Phosphorus movement in a Boreal Plain soil (Gray Luvisolic) after forest harvest

By

Ivan Richard Whitson ©

A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfillment of the requirements for the degree of Doctor of Philosophy

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The undersigned certify that they have read, and recommended to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled *Phosphorus movement in a Boreal Plain soil (Gray Luvisolic) after forest harvest* submitted by Ivan Richard Whitson in partial fulfillment of the requirements for the degree of Doctor of Philosophy in Environmental Biology and Ecology.

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**Dedication**

To Jonah Robert Douglas, Alana Regan, and Sawyer Vaughan, who made this venture far more interesting and rewarding.

**Abstract**

Understanding the effect of forest harvest on phosphorus export from the hillslope requires an appreciation of the soil profile's influence on processes that control phosphorus movement. Water movement patterns and phosphorus concentration trends were studied in Gray Luvisolic soil-dominated hillslopes to infer the effect of forest harvest on phosphorus movement in typical Boreal Plain uplands. I hypothesized that phosphorus concentrations would decrease with soil depth, and that interflow would occur through upper horizons. Phosphorus movement was expected to increase after forest harvest due to higher phosphorus concentrations and greater interflow. Hydraulic properties were determined by measurements of interflow, bromide movement, soil water content, infiltration, and hydraulic conductivity. Phosphorus concentration and retention were determined with *in situ* soil water and air-dried samples of soil horizons. The probable solid phases controlling orthophosphate solubility were identified after modeling phosphorus complex formation.

The minimal interflow observed was partially due to precipitation of just 50% of normal and subsequent dry soils. However, reduced infiltration capacity and equal vertical hydraulic conductivity of Ae and Bt horizons at harvested sites suggested that harvest also diminished interflow. Despite these hydraulic changes, most snowmelt infiltrated while soils remained at 0 °C. Most applied bromide remained above 60 cm depth, and more was recovered at forested than harvested sites 13 months after application. Loss of bromide reflected vegetation uptake and lateral convective-diffusive movement. Mean soluble reactive phosphorus concentration decreased with depth, ranging from 64 mg $L^{-1}$ in the LFH horizon to 0.01 mg $L^{-1}$ in the groundwater zone; orthophosphate was the dominant phosphorus

complex in the soil solution. The solid-phase phosphate minerals in equilibrium with orthophosphate ranged from secondary Fe-, and Ca-phosphates in the LFH to fluoroapatite in the lower Ae. Adsorption reactions probably controlled orthophosphate activity in subsoil horizons. Both extractable and water soluble phosphorus decreased while pH increased after harvest in the LFH horizon only. Downslope phosphorus movement is potentially greatest when soils are moist and frozen. However, hydraulic changes and reduced phosphorus concentration that resulted from harvest reduced the potential for downslope phosphorus movement under the conditions of this study.

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# Chapter 1

## The development and testing of hypotheses concerning phosphorus movement in a Boreal Plain soil

### *Movement of Phosphorus and Water*

Does phosphorus move in the landscape? If so, why, and how does it travel, and why would it matter? These are the questions I set out to answer when I first began graduate research. While some of the literature suggested that in most cases, phosphorus was immobile (Peaslee and Phillips 1981), I began to find support for the alternative view, and good reasons to pursue the research. Phosphorus has critical implications for the health of our surface water resources. An agricultural innovation of the 19$^{th}$ century was the recognition of the importance of phosphorus in terrestrial ecosystems (Brady 1974). Our understanding of the role of phosphorus in aquatic systems came more recently with the finding that phosphorus fertilization elevated algal productivity and caused eutrophication in nutrient-poor (oligotrophic) lakes on the Canadian shield of NW Ontario (Schindler *et al.* 1971). Phosphorus was later found to limit aquatic productivity in western Canadian lakes (Prepas and Trew 1983), many of which have an internal supply of phosphorus that re-circulates between the water column and the sediments during anoxic conditions (Prepas and Vickery 1984; Riley and Prepas 1984). If phosphorus is mobile within watersheds, increased inputs into western Canadian lakes may threaten many eutrophic lakes with declines in dissolved oxygen (Babin and Prepas 1985) and production of blue-green algae and associated toxins (Trimbee and Prepas 1987; Kotak *et al.* 2000; Prepas *et al.* 2001a). Hence, movement of phosphorus is a compelling issue in this region.

But does phosphorus actually move in the landscape? One way to answer this question is to examine the record stored in soil profiles, with particular reference to those in the Boreal Plain ecozone of western Canada. There is an abundant supply of apatite-phosphorus in many Western Canadian mineral soils inherited from the underlying marine and non-marine sedimentary bedrock. The chemical environment associated with these geologic materials was conducive to apatite formation and stability (Stumm and Morgan 1981, p. 284). Apatite, however, is unstable under acidic conditions that accompany soil genesis under forest vegetation. In northern Saskatchewan for instance, approximately 40% of phosphorus originally present in the top 50 cm of Gray Luvisolic soils was removed over a 10,000 year period, possibly by leaching (Xiao *et al.* 1991). Studies in other regions show that leaching rates are greatest in younger soils (Walker and Syers 1976) when primary calcium-phosphates like apatite are converted to secondary or organic forms. Complete loss of primary apatite occurred in New Zealand soils after 10,000 to 70,000 years of development (Walker and Syers 1976). The phosphorus in older, more acidic and weathered soils is held as stable iron-, aluminum-, and organic phosphorus compounds that are of very low solubilility (Smeck 1985). Conversion of calcium-phosphates to these other forms has been documented on the Boreal Plain in both fine-textured (Frossard *et al.* 1989; Schoenau *et al.* 1989) and coarse-textured soils (Fyles and McGill 1988), as are gradual increases in aluminum- and iron-oxide contents of some in the more acidic soil horizons in the region (Pawluk and Lindsay 1964; St. Arnaud and Whiteside 1964). The rapid loss of phosphorus from forest soil profiles in the region (Frossard *et al.* 1989) has been explained as being due to the greater mobility of organic forms of phosphorus (Frossard *et al.* 1989; Donald *et al.* 1993) or due to lateral subsurface flow above fine-textured subsoil horizons (Xiao *et al.* 1991). Peatlands, second in areal extent to Gray Luvisols on the Boreal Plain (Ecological Monitoring and Assessment Network 1996) experience very slow organic phosphorus mineralization rates because of anoxic conditions and low temperatures (Mugasha *et al.* 1993) and represent a part of the landscape where phosphorus accumulates in stored organic matter. The dominance of Gray Luvisolic soils on the Boreal Plain ecozone (Ecological Monitoring and Assessment Network 1996) suggests that phosphorus movement in the terrestrial ecosystem has been widespread. The phosphorus stored in peatlands may have been derived from weathering and subsequent leaching of Gray Luvisols. Transport of particulate phosphorus through the atmosphere should not be ignored (Shaw *et al.* 1989) as a possible mechanism of removal either, particularly since forest fires occur so frequently on the Boreal Plain. But if phosphorus losses inferred from soil genesis studies are related to water movement, then an examination of hillslope runoff processes will help understand the phosphorus leaching patterns of today.

It is unlikely that phosphorus in streams is derived equally from all parts of the watershed. Sources of runoff are highly variable in watersheds (Freeze 1972; Anderson and Burt 1990). Solutes like dissolved organic carbon are often derived principally from high supply areas within the watershed such as the riparian areas, that in Boreal regions often include wetlands (Dosskey and Bertsch 1994). Correlations between peatland area in watersheds and phosphorus concentration in streams and lakes for both the Boreal Shield (Dillon and Molot

1997) and Boreal Plain (Devito *et al.* 2000; Prepas *et al.* 2001b) ecozones suggest that losses from Luvisols may, for instance, simply have been a transfer within the terrestrial system towards fen wetlands (Vitt *et al.* 1995). Strong differences in phosphorus retention properties between soils of the two regions, however, suggest that uplands of the Boreal Plain ecozone may play a much different role than similar sites on the Boreal Shield in phosphorus transport to water bodies.

Can the insights of long-term phosphorus movement patterns on the Boreal Plain ecozone help us to predict the implications of forest harvest over shorter time scales? Yes, but only after a digression into an analysis of the factors that are likely to control water movement and how these are affected by forest removal. Water movement in landscapes is spatially and temporally variable, affected by the geometry of soil pores, the layering of profiles, the topography of the land surface, and climate. Several important runoff mechanisms observed in other regions are likely to be found on the Boreal plain ecozone. Infiltration excess-overland flow (Horton 1933) has been reported for tropical forests (Wierda *et al.* 1989; Bonell 1993; Elsenbeer and Lack 1996) but rarely in temperate forest soils (Dunne 1978; Buttle and Turcotte 1999) except in unusual circumstances such as the case of exposed bedrock (Allan *et al.* 1993) or when ice layers exist in snow (Price and Hendrie 1983). This runoff type is expected only in the case of intense soil disturbance because infiltration capacities of upper horizons are typically very high. Saturated excess overland flow (Dunne and Black 1970) is most likely to be associated with areas of high water tables such as peatlands and soils of the Organic (Devito *et al.* 1996) and Gleysolic (Taylor 1982) Orders.

Interflow, the infiltration of water and subsequent lateral downslope movement through upper soil horizons (Gray and Wigham 1970) typically results from the presence of flow-restricting layers (Whipkey 1965; Whipkey and Kirkby 1978; Ahuja *et al.* 1981). Interflow has been widely reported in regions with dissimilar soil characteristics to those of the Boreal Plain, such as soils with shallow bedrock in Britain (Weyman 1973), soils formed on conglomerates in New Zealand (Mosley 1979; Mosley 1982; McDonnell 1990), volcanic soils in the NW USA (Harr 1977) and on the Canadian Shield (Renzetti *et al.* 1992; Peters *et al.* 1995). Interflow has also been found in landscapes where soil hydraulic properties are similar to Gray Luvisols, such as on semi-arid, high elevation pine forest in New Mexico (Newman *et al.* 1998), on humid Ultisolic soil landscapes of the SE USA (Gaskin *et al.* 1989; Mulholland *et al.* 1990; Wilson *et al.* 1990; Dosskey and Bertsch 1994), and on drumlins in Ontario (Buttle 1989). Complex soil stratigraphy of the Boreal Plain is partially due to the great depth (Pawlowicz and Fenton 1995) and varied genesis (Pawluk and Bayrock 1969; Shaw 1994) of glacial deposits such that flow restricting layers are likely to be common in any vertical sequence of regolith. Transport of clay to subsoils, the dominant genetic process in Gray Luvisol formation (Howitt and Pawluk 1985a; Howitt and Pawluk 1985b), has contributed to a general pattern of reduced hydraulic conductivity at shallow depths in the soil profile (Coen and Wang 1989). Hillslopes in the region then have the characteristics that allow interflow. With overland flow situations, particulate phosphorus is likely to dominate, whereas with interflow, soluble phosphorus is likely to be more important. Phosphorus concentration in overland flow will depend on fluid velocity and substrate erodibility, whereas mineralogical properties will determine the solubility of phosphorus along the interflow pathway.

In Chapter 2, I hypothesized that interflow would be an important mechanism transmitting flow in Gray Luvisols. I used simple flow interception techniques (Atkinson 1978) to investigate saturated water movement in upper soil horizons and used a conservative tracer, calcium bromide, to assess movement during both saturated and unsaturated conditions. Tracers have been used to assess the convective and dispersive components of downward water movement in the soil profile (Butters and Jury 1989; Nachabe *et al.* 1999). They have also been used to trace lateral water movement in soil profiles, including highly weathered tropical soils (Agus *et al.* 1998), fine-textured temperate region forest soils (Bruce *et al.* 1985; Steenhuis and Muck 1988; Bathke *et al.* 1992; Wilson *et al.* 1993; Afyuni *et al.* 1994), coarse-textured groundwater aquifers (Mackay *et al.* 1986), and loess-derived soils (Reuter *et al.* 1998). Tracer studies have shown that soil horizon boundaries limit downward flow (Newman *et al.* 1997) and have been used to assess the importance of flow in different pore-size domains of the soil (Bowman and Rice 1986; Jardine *et al.* 1990; Tsuboyama *et al.* 1994; Bronswijk *et al.* 1995).

When water flows through a soil in accordance with Darcy's law, it moves from a position of higher potential energy to lower in proportion to the resistance of the media, or hydraulic conductivity, through which flow takes place. The variation in hydraulic conductivity spatially (e.g., depth) or temporally (e.g., after logging) then controls the possible occurrence of both interflow and overland flow. Interflow is favored at locations in the profile where lateral hydraulic conductivity exceeds vertical (Bathke and Cassel 1991). Hydraulic conductivity is a function of porosity, soil texture, soil structure (Bouma 1982; Southard and Buol 1988), and the presence of cemented layers (De Kimpe and McKeague 1974; Mehuys and De Kimpe 1976; Dabney and Selim 1987; Habecker *et al.* 1990; Day *et al.* 1998). Structural effects on hydraulic conductivity are often more important than texture or bulk density (Schoeneberger *et al.* 1995) and may be partially inherited from the parent material (Schoeneberger and Amoozegar 1990). Hydraulic conductivity is highly variable (Sharma *et al.* 1980), often lognormally distributed especially under unsaturated conditions (Warrick and Nielsen 1980; Mecke *et al.* 2000), and is affected by measurement method (Mohanty *et al.* 1994). A lognormally distributed variable must be transformed to reduce the variance and create the normal distribution that allows parametric testing. Measurement is problematic in soils with macropores (Bouma 1982; Bouma 1983; Douglas 1986; Buttle and House 1997) or in those soils derived from unsorted parent materials like till (Espeby 1990a). Further complications to this parameter result from swelling behavior in clay-dominated soils (Garnier *et al.* 1998; Lin *et al.* 1998) such that measurements are influenced by initial water content. In spite of these challenges, measurements of hydraulic conductivity are a common basis for predicting water movement patterns.

Overland flow occurs when rainfall intensity exceeds soil infiltration rates, but infiltration capacity in forest soils is very high, especially in litter layers (Tsukamoto 1975; Dunne 1978). Soil disturbance associated with logging machinery and tree hauling can reduce infiltration rates through increases in bulk density (Corns 1988; Cullen *et al.* 1991; Stone and Elioff 1998; Dykstra and Curran 2000; Whitson *et al.*, 2003) and increase the importance of overland flow (Heede 1987; Standish *et al.* 1988). Increased overland flow will mean an increased transport of phosphorus in entrained sediment (Abrams and Jarrell 1995). In Chapter 3, infiltration rates and hydraulic conductivity are explored for both harvested and undisturbed forest conditions. I hypothesized that the hydraulic conductivity of the Bt horizon of Gray Luvisolic soil would be less than that of the Ae horizon. I also hypothesized that forest harvest would increase bulk density and reduce infiltration capacity. Hydraulic conductivity was measured with undisturbed soil cores while infiltration was measured with double-ring infiltrometers.

***Seasonal Influences***

Water movement during the winter months in the Boreal region will differ from that of summer because of the effects of freezing on soil hydraulic properties (Wilcox *et al.* 1997). Snowmelt is often a major component of annual runoff in Boreal (Martz 1978; Maule and Stein 1990) and Arctic streams or lakes (Kane *et al.* 1991) and is important for groundwater recharge on the western Canadian Prairie ecozone (Maule *et al.* 1994). Simulation of water movement under frozen conditions is made difficult by the changes to effective soil porosity and hydraulic conductivity that result from freezing (Barry *et al.* 1990; Prevost *et al.* 1990; Stahli *et al.* 1996). Frozen soil reduces infiltration rate (Gray *et al.* 1970; Dunne and Black 1971; Rybakova 1990), with the effect of frost found to be inversely-related to soil water content at the time of melt (Granger *et al.* 1984). Soil freezing resulted in ice lens formation in silt textured soil in Alaska (Kane and Stein 1983) and in Gray Luvisolic Ae horizons in Alberta (Howitt and Pawluk 1985b). Differences in relative soil moisture saturation may explain some of the differences between runoff behavior in areas of continuous (Marsh and Woo 1984), and seasonal frost. Also, there was less snowmelt runoff on seasonally-frozen soils with a southerly aspect than on a permanently frozen north-facing slope in the Yukon (Carey and Woo 1998). Solute transport is influenced by the presence of ice-blocked soil pores, with preferential flow observed (Nyberg and Fahey 1988; Espeby 1990b). Chapter 4 of the thesis examines a hillslope during the snowmelt period. I hypothesized that frozen soil would generate subsurface flow through both the Ae and LFH horizons. The timing, path and flux of water were compared to soil temperature and moisture changes during the period of snowmelt and soil thaw.

***Adding Phosphorus to the Equation***

Understanding phosphorus movement requires, in addition to the nature of water movement processes, an appreciation of the factors that control the amount of phosphorus in water. One reason for the perceived immobility of phosphorus is the typically low dissolved concentrations in water (Clayton and Kennedy 1985; Martin and Harr 1989). For this reason, particulate phosphorus is often the major form that moves (Meyer and Likens 1979; Vaithiyanathan and Correll 1992), including on the Boreal Plain (Munn and Prepas 1986; Cooke and Prepas 1998).

The relationship between solution and solid phase phosphorus can be approached with the principles of thermodynamic equilibrium applied by Lindsay and others to the soil environment (Lindsay and Moreno 1960; Lindsay 1979; Lindsay 1981; Keizer 1996). The activity of inorganic phosphorus (orthophosphate) is maintained by the solubility product of various metal-phosphate species. When the ion-activity product of any metal-phosphate species exceeds its solubility product, precipitation can occur. Reactions that control the activity of metals, anions, or other agents like pH and carbon dioxide then will also influence phosphorus solubility. In acidic soils, aluminum and iron phosphates generally dominate, whereas in soils of neutral pH, calcium phosphates are generally formed (Smillie *et al.* 1987; Syers and Curtin 1989; Cho 1991). The success of applying solubility theory depends on the accuracy of the thermodynamic equilibrium constants and the correct identification of components of the soil system. The approach can also be frustrated because soil solutions often reach equilibrium very slowly. Other faster reactions such as biotic uptake or adsorption may be more important in controlling phosphorus concentration over short time scales. Nevertheless, this technique is likely best applied to forest systems where the phosphorus cycling process is largely closed, at least relative to agricultural systems or riparian areas, where phosphorus gains and losses perturb the equilibrium.

The relation between solution and solid-phase phosphorus can be described equally well using the concept of adsorption, the accumulation of a substance at a phase boundary like the solid-liquid interface (Sposito 1984). While adsorption in practice is difficult to distinguish from chemical precipitation (Hsu and Rennie 1962; Sposito 1984) it is still a useful approach to understanding the disappearance of a solute from solution (Enfield and Ellis 1983). The functional relationship between the amount of phosphorus in solution and the solid phase at equilibrium obtained under standard conditions of temperature, pressure, and background electrolyte, is termed an adsorption isotherm. Isotherms have been described by a variety of equations (Parfitt 1978; Sibbesen 1981), including the purely empirical Freundlich equation (Fitter and Sutton 1975) and the physically-based Langmuir equation (Harter and Smith 1981). If its underlying assumptions are met, the Langmuir equation yields information on the maximum density of adsorbed phosphorus and also a measure of the average bonding energy at phosphorus sorption sites. Adsorption studies are commonly employed to infer the relationship between the solid phase and soil solution phosphorus concentrations over short time scales.

Depending on the soil, adsorption parameters correlate with the content of crystalline or amorphous iron- and aluminum-oxides (Ryden and Pratt 1980; Laverdiere and Karam 1984; Johnson *et al.* 1986; Zeng *et al.* 2002), extractable aluminum and organic matter (Saini and MacLean 1965; Soon 1991), extractable iron (Scheinost and Schwertmann 1995), clay content (Olsen and Watanabe 1957; Soon 1991; Sanyal *et al.* 1993; Singh *et al.* 1996), and calcium carbonate (Weir and Soper 1962). Adsorption has in some situations been explained as the exchange of phosphate for an OH group found at a surface $Al\text{-}(OH)_2$ site (Muljadi *et al.* 1966a; Muljadi *et al.* 1966b; Parfitt 1978). Adsorption capacity depends on the density of reactive OH species, with acidic soils generally having higher density for phosphorus adsorption (Olsen and Watanabe 1957). Iron- and aluminum-oxides and aluminosilicate minerals like montmorillonite and kaolinite possess these $Al\text{-}(OH)_2$ groups but at differing density and bonding energies for phosphorus sorption. Soils derived from volcanic materials also have high sorption capacities (Imai *et al.* 1981; Abrams and Jarrell 1995; Zeng *et al.* 2002) because of abundant oxides. In aluminosilicate clay minerals, such groups are typically limited to mineral edges (Pissarides *et al.* 1968; Beek and van Riemsdijk 1982). Adsorption also occurs on calcite surfaces (Freeman and Rowell 1981) or on organic matter (Rennie and McKercher 1959; Zhou *et al.* 1997). Crystalline aluminum and iron oxides adsorb 5 to10 times more phosphorus per unit basis than aluminosilicates or carbonates, while amorphous iron and aluminum oxides adsorb about 1000 times more phosphorus per unit area than aluminosilicate minerals (Ryden and Pratt 1980). Adsorption reactions can be very effective, particularly in weathered soils, at keeping solution phosphorus concentration low (Riekerk 1971; Wood *et al.* 1984; Yanai 1991). Anoxic conditions generally reduce sorption capacity (Nair *et al.* 1999) due to greater solubility of iron oxides in the reduced state. Soil types that are associated with poor drainage will retain less phosphorus than those that tend to have oxic conditions.

The time of contact between phosphorus in solution and the solid phase is important to controlling phosphorus movement (Sharpley and Syers 1979). Adsorption and desorption reactions occur relatively quickly as compared to the slower precipitation reactions (Parfitt 1978), nonetheless, they may still operate too slowly for dissolved phosphorus in moving water to reach an equilibrium. Reaction rates become more important when solute movement occurs rapidly (e.g., preferential flow) as opposed to slowly (e.g., displacement flow) relative to the solid phase (Buttle and Sami 1990). Another complication of adsorption reactions is the hysteresis observed for desorption. That is, the surface excess at any solution concentration is usually greater when desorption is occurring. Desorption reactions are partially irreversible, with some phosphorus no longer available for re-entering solution

(Sanyal *et al.* 1993). The irreversibility is attributed to chemical precipitation, surface penetration or diffusion into the interior of the soil mineral (Ryden and Pratt 1980). This slow change in surface reactivity means that the adsorption sites gradually regenerate and can adsorb new phosphorus again as older phosphorus diffuses into the solid's structure (Ryden and Pratt 1980).

The regions of the soil through which water moves has implications for solute flux (Luxmoore *et al.* 1990; Wilson *et al.* 1991), and can be expected to control the magnitude and proportion of soluble and particulate forms of phosphorus. Water that travels through the soil shows varying degrees of mixing between pore regions (McDonnell 1990; Turton *et al.* 1995). For instance, the amount of aluminum transported in interflow varied between micropore and macropore regions and by soil depth (Cozzarelli *et al.* 1987; Hendershot *et al.* 1992). Dissolved organic carbon and calcium transport varied with flowpath depth (McGlynn *et al.* 1999). Inorganic phosphorus transport has been facilitated by macropore flow in clay textured soils (Djodjic *et al.* 1999; Jensen *et al.* 1999). Overland flow is capable of transporting particulate phosphorus in proportion to sediment load while phosphorus transport in interflow is limited by the solubility and adsorption characteristics encountered along the flowpath.

Interactions between phosphorus and dissolved organic carbon can affect movement in several additional ways. Organic acids like fulvic acid abound in the forest floor (Beyer *et al.* 1993) and so are abundant in forest systems. Dissolved organic phosphorus may be adsorbed less strongly than soluble inorganic forms (Frossard *et al.* 1989; Donald *et al.* 1993). Phosphorus sorption is reduced in the presence of some organic acids (Traina *et al.* 1986; Ohno and Erich 1997). Low molecular weight organic acids produced by decomposition of forest litter can displace phosphorus from adsorption sites (Fox and Comerford 1990; Fox 1993). Inorganic phosphate can also exist as complexes with metals and organic acids (Schnitzer 1969; Arp and Meyer 1985; Pohlman and McColl 1988). Not surprisingly then, organic, as opposed to inorganic, dissolved phosphorus is often reported in forest systems (Timmons *et al.* 1977; Yavitt and Fahey 1986; Qualls and Haines 1991; Qualls *et al.* 1991; Cortina *et al.* 1995; Correll *et al.* 1999). Likewise, the downward transport of phosphorus observed in forested Luvisolic soils in Saskatchewan (Schoenau and Bettany 1987) has been explained as a case of phosphate-fulvic acid complexation. The body of theory governing inorganic phosphorus reactions may not apply well to forest systems.

My objective in Chapter 5 was to examine phosphorus solubility and retention properties in Gray Luvisolic soil profiles. I hypothesized that upper horizons such as the Ae would retain less phosphorus than subsoil, and that solution concentrations would decrease with depth. Differences in phosphorus affinity and solution concentration among horizons were used to draw conclusions about how different parts of the soil profile would influence dissolved phosphorus concentration in moving water.

***Forest Disturbance***

Imposition of forest harvesting in other regions has lead to increases in runoff and solute flux (Bormann *et al.* 1968). Forest harvesting has reduced transpiration and increased annual water yield of catchments (Hibbert 1965; Hewlett and Helvey 1970; Hornbeck *et al.* 1993; Verry 2000), increased deep percolation rates (Gessel and Cole 1965), or shifted the hydrograph forward in time (Wright *et al.* 1990). Forest harvesting is expected to accelerate phosphorus export in the Boreal Plain ecozone through changes to runoff and dissolved phosphorus concentration. Kachanoski and De Jong (1982) found that hillslope water yield increased by five times during the snowmelt period and doubled during the growing season at a site in northern Saskatchewan. Forest harvesting resulted in increased total phosphorus concentration in Boreal Plain lakes during the year following harvest (high precipitation) but not in the second post-harvest year (low precipitation) (Prepas *et al.* 2001a). Changes to runoff are a reflection of reduced evapotranspiration, with subsequently wetter soils (Elliott *et al.* 1998). While the hydrologic effects of forest fire differ somewhat from harvest (Chanasyk *et al.*, 2003), they are likely to differ strongly in their effect on phosphorus in near-surface horizons (Romanya *et al.* 1994; Malmer 1996). Changes in solar energy inputs (Standish *et al.* 1988; Ballard 2000) are likely to result in higher carbon decomposition and phosphorus mineralization rates (Gressel *et al.* 1996) such that solution phosphorus concentrations increase after logging. However, change in species composition, towards grass for instance, can reduce energy inputs and soil temperature in Boreal soils (Hogg and Lieffers 1991). Changes in species composition towards those with higher evapotranspiration rates or greater potential for interception resulted in reduced annual runoff (Hornbeck *et al.* 1993). Changes to energy inputs have resulted in earlier snowmelts (Berry and Rothwell 1992; Meng *et al.* 1995). Forest harvest is expected to increase phosphorus export rates until such time that runoff rate and solution phosphorus concentration return to predisturbance values. Disturbance effects themselves may be obscured or exacerbated by climatic trends (Paterson *et al.* 1998; Schindler 2001).

The objective of Chapter 6 was to infer differences in potential phosphorus flux due to forest harvest. I hypothesized that logging would make soils wetter, thaw earlier, and have more subsurface flow. I also expected that extractable phosphorus would increase after harvest. A variety of hillslope sites with differing topography were compared through measurements of soluble phosphorus concentration, extractable phosphorus, bromide transport, soil temperature, and moisture content.

The implications of forest disturbance on phosphorus movement in Gray Luvisols then will depend on how much runoff derived from upper horizons increases and what changes occur to the abundance of phosphorus in solution. Changes to the relative importance of different hydrologic flowpaths are critical. Increased flow through the LFH horizon would greatly increase the amount of phosphorus moving to lower slope positions. Although the amount of flow, closely related to precipitation, probably determines long-term average rates of downslope movement, the chemical changes with depth are probably equally important, simply because phosphorus concentrations vary over several orders of magnitude with soil depth. If soil solution phosphorus concentration is unaffected by forest removal, phosphorus export should increase in proportion to runoff. However, results from the TROLS study (Prepas *et al.* 2001a) showed that harvesting increased total phosphorus content of study lakes in a wet year with higher runoff but not in a dry year of low runoff. In Chapter 7, the observations made in earlier chapters are summarized to develop new hypotheses regarding how phosphorus and water movement patterns that apply over short time scales are linked to the long-term evidence and to the larger landscape as a whole.

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## Chapter 2

## Hydrologic flowpaths in a recently-logged Western Canadian forest soil

### *Introduction*

Forest removal is likely to impact water movement patterns on hillslopes within the Boreal Plain Ecozone of western Canada. While logging has been shown to increase runoff from plots via surface and subsurface flowpaths in sandier upland soils in the region (Kachanoski and De Jong 1982), there has been limited study of water movement on hillslopes with typical fine textured soils. The contribution of individual hydrologic flowpaths towards runoff becomes especially important when solute mobility is considered. Mobility of aluminum, for instance, was related to the hydrologic flowpath (Cozzarelli *et al.* 1987) and the same is likely to apply to phosphorus, a macronutrient extremely important to regional water quality through its influence on productivity of regional lakes (Prepas *et al.* 2001). A change towards shallower flowpaths will mean an increase in the phosphorus export rate in this region (Evans *et al.* 2000). Thus, potentially increased runoff after forest removal has implications for both quantity and quality of water downslope of disturbed areas.

Forest removal can severely disturb upper soil horizons and alter flow patterns. Logging for instance has been associated with increased bulk density and surface crusting, (Cullen *et al.*

1991; Ballard 2000) although the magnitude of its effect depends on soil texture, soil moisture content, season of disturbance and extent of post-harvest scarification. While logging may enhance overland flow conditions through its negative effect on infiltration capacity, its effect on interflow is uncertain. At larger scales, effects of local impacts on either the watershed hydrograph or solute transport may disappear due to dilution (Hornbeck *et al.* 1993).

Interflow has been defined as a process where water that has infiltrated the land surface moves laterally downslope through upper soil horizons until it emerges at the land surface again (Gray and Wigham 1970, pg. VII.4). A reduction in vertical hydraulic conductivity ($K_v$) with depth in the profile is a requirement for interflow (Whipkey 1965). However, $K_v$ reduction alone does not adequately explain the widely varying location of interflow in profiles similar to Gray Luvisols at various North American sites. In Ultisolic soil landscapes of the humid eastern US, interflow was associated with clay-dominated Bt subsoil horizons (Gaskin *et al.* 1989; Mulholland *et al.* 1990; Wilson *et al.* 1990), and not above in the coarser textured A horizons. Ultisols and Luvisols possess clay-dominated subsoil horizons, but differ in terms of chemical properties (Miller 1983; Howitt and Pawluk 1985a; Howitt and Pawluk 1985b). In the semi-arid south-western US, interflow occurred within a clay-dominated Bt subsoil horizon of a Luvisol equivalent that had developed from weathered volcanic-derived bedrock material, (Wilcox *et al.* 1997) and not below in unweathered soil parent material of much higher $K_V$. The location of interflow in these examples is not necessarily through horizons of higher $K_{sat}$.

The actual location of interflow in the soil profile is best predicted by reference to vertical discontinuities in soil structure, texture, bulk density or porosity that ultimately affect the relative magnitude of the $K_v$-$K_l$ (vertical-lateral hydraulic conductivity) ratio at a particular depth. Any location in the profile where $K_l$ exceeds $K_v$ can potentially experience interflow. Interflow is likely to be an important flow process on hillslopes on the Boreal Plain, where Gray Luvisolic soils developed on clay-dominated glacial deposits dominate (Marshall and Schut 1999). Two locations in the Gray Luvisolic profile are likely to be important for interflow production: 1) the organic/mineral interface, typically an LFH /Ae horizon sequence where the litter layer overlies mineral soil, and 2) the Ae/Bt horizon interface, where the relatively coarse textured Ae horizon overlies the clay-dominated Bt horizon. These typically occur at depths < 30 cm. The LFH horizon reflects litterfall accumulation that has occurred between fire intervals and typically has a bulk density of 0.1-0.2 Mg $m^{-3}$. Increased clay content of the Bt compared to the Ae horizon means a decrease in average pore size across the interface between these two horizons. Soil structure, or the tendency to form aggregates of a particular geometric shape, changes across that interface as well. The structural units of the Ae horizon break into horizontal platelets, whereas the Bt horizon typically has a subangular blocky structure (Howitt and Pawluk 1985a). The size and orientation of these interplanar voids will greatly influence the $K_v$-$K_l$ ratio. The characteristic platy structure of Ae horizons reflects episodes of freezing and thawing over long periods (St. Arnaud and Whiteside 1964) and is expressed more strongly in Luvisols of colder climates (Rust 1983). These structural differences probably impart anisotropic conditions to the Ae horizon, such that $K_l$ exceeds $K_v$ within the horizon itself (Whitson *et al.* 2003b). Structural and textural characteristics in the upper profile suggest that interflow will occur frequently in Gray Luvisolic soils.

***Objectives and Hypotheses***

This study's objective was to determine the water movement patterns along near-surface hydrologic flowpaths on a recently harvested hillslope with Gray Luvisolic soil. I hypothesized that:

- The infiltration capacity of Gray Luvisolic soils exceed rainfall intensities by a wide margin such that overland flow does not occur in forested or logged situations.
- The fine textured subsoil of Gray Luvisolic soil profiles allows downward percolation of moisture.
- The fine-textured Bt horizon immediately beneath the relatively coarse textured Ae horizon of the Gray Luvisolic soil profile promotes interflow under non-frozen conditions towards lower slope positions along the interface of the two horizons.

Interflow and percolation were thus expected to be the dominant hydrologic flowpaths in this logged setting.

***Site Description and Methodology***

The site was located within the Mid Boreal Mixedwood Ecoregion, (Strong and Leggat 1992) approximately 250 km north of Edmonton, Alberta, Canada. Soils of the Luvisolic order dominate upland areas, whereas peatlands (Organic Order and Gleysolic soil Orders) dominate poorly drained depressional areas (Wynnyk *et al.* 1963; Strong and Leggat 1992). Vegetation of upland locations reflects local fire history, ranging from aspen (*Populus tremuloides*) to white-spruce (*Picea glauca*) dominated. Other important tree species include *P. balsamifera* (balsam poplar), *Pinus banksiana* (jack pine), *Betula papyrifera* (birch) and *Abies balsamea* (balsam fir). Glacial drift in the area is thicker than 15 m (Pawlowicz and Fenton 1995) and overlies sedimentary bedrock of Upper Cretaceous age (Green 1972). The study area had an elevation of approximately 685 m above sea level and was located within the Driftwood River catchment, a 2100-km$^2$ basin that drains the southern slopes of the Pelican Mountains into the Athabasca River near Smith, Alberta. Mean annual precipitation for the period 1970-90 at nearby Slave Lake, Alberta was 496 mm, of which 30% fell as snow (Environment Canada 2000). This location was used for climate data even though it lies outside the watershed because there was no similar record of climatic data from within the Driftwood catchment. The Slave Lake climate station occurs at similar elevation and possesses similar climatic conditions as within the watershed. Environment Canada's monitoring of flow indicated that catchment runoff averaged 119 mm per year for the period 1970-90, with 28 and 56 % of annual runoff occurring for the periods April-May and June-August respectively (Environment Canada 2002). This meant that stream runoff was probably rainfall, not snowmelt, dominated for these 2 decades. Using mean potential evapotranspiration for the region, moisture deficits often exceed 100 mm per year (Dzikowski and Heywood 1990; Strong and Leggat 1992).

The hillslope chosen for the study had been recently harvested and had a uniform soil and topography typical of the hummocky morainal landscapes of the Plains region. A convex hillslope segment with an 11-13% gradient at midslope and easterly aspect was selected for study (Figure 1). Mature aspen and white spruce were harvested during the winter of 1996-97, but a narrow buffer was left along the lower slope to protect an adjacent lake. Harvested blocks of upland forest ranged in size from 10 to 40 ha. Observations at the time of harvest suggest that soils were frozen but the depth of frost was unknown. Snow depth was approximately 40 cm. Logging was carried out by a combination of cut-to-length and whole-tree-length techniques. With the former, trees were processed at the stump with a Timber Jack 1270 and wood was hauled to a logging road with a Timber Jack 1210 Forwarder. With whole-tree-length harvest, a Timber Jack 618 Feller-Buncher cut the trees and a grapple skidder moved the logs to roadside for de-limbing and subsequent loading. Harvested areas received no further post-harvest silvicultural treatment other than the planting of white spruce seedlings. No silvicultural treatments were employed on the harvested area, except that part of the hillslope was replanted with white spruce seedlings in 1997 near the time my study began. Aspen suckers and other forest plants re-vegetated the hillslope and surrounding area the summer following harvest. Instrumentation was installed in 1998 and sampling conducted along three transects (1-3) perpendicular to the slope at up to four slope positions ranging from lower to upper. Transects were used to capture the expected high spatial variability within each slope position. Three transects were chosen as a balance between affordability and scientific validity. Transect 1 occupied the northern edge of the hillslope segment, Transect 2 the centre and Transect 3 the southern edge (Figure 1.). Soil profiles were examined at each slope position and sampled for physical and chemical properties. Soil particle size was determined by the hydrometer method (Sheldrick and Wang 1993) and organic carbon content by dry combustion (Tiessen and Moir 1993). Depth to free calcium carbonate was considered to mark the beginning of unweathered soil parent material. Bulk density was measured on two occasions over the period of the study. The first set of measurements were obtained at soil pits dug at the upper midslope position of each transect and at one additional soil pit approximately 1-km distant the first summer after logging. The second set of bulk density measurements were taken at 1 location on the hillslope and at 8 other locations with similar soils within a 1-km radius of the hillslope the third summer after harvest. Samples were collected by Uhland core (7.6 cm diameter by 7.6 cm length) in increments to 160-cm depth at up to six replicates per increment for each location. High expected spatial variability lead us to use a large sample size for bulk density. Porosity was calculated from measured bulk density assuming a mineral soil horizon particle density of 2.65 Mg $m^{-3}$ and an organic soil horizon particle density of 1.00 Mg $m^{-3}$. Saturated hydraulic conductivity was measured by falling head permeameter (Reynolds 1993) with undisturbed soil cores taken from 6 Gray Luvisolic soil profiles, 4 from this location and 2 from another location 100 km distant.

Precipitation at the study site was estimated from a combination of on-site measurements augmented by data from nearby weather monitoring stations. Precipitation was measured 1.0 km distant from the hillslope for June-October of 1997 with a tipping bucket rain gauge and datalogger. In 1998 and 1999 precipitation for the June-October period was measured on site with a manual rain gauge, checked approximately weekly. Missing data and data for

May of all three years were estimated from two weather stations at forest fire towers located about 20 km northwest and southeast, respectively from the site. Snowpack water equivalent was determined on the hillslope from measurements of depth and density in late February of both 1998 and 1999.

Calcium bromide was used to trace soil water movement. Loss of bromide from a sample volume of soil was assumed to result from interflow, providing the bromide plume remained above the maximum depth of sampling and providing that plant uptake of bromide was prevented (Figure 2). Plots were established at the upper midslope position of each transect where the slope gradient was very similar (Figure 1). Tracer was applied on September 26-27, 1997. The litter layer (LFH horizon) was temporarily removed from a 4-$m^2$ square area to allow application of bromide to the mineral soil (Ae horizon) surface, but was replaced immediately so that infiltration would not be affected afterwards. Roots associated with the LFH and upper Ae horizons were removed during this process resulting in no aspen regrowth on the plots for the remainder of the study. Vegetation removal was necessary to allow all bromide to enter the mineral soil at the time of application, although it was understood that removal would influence soil water movement patterns. Application of bromide to the vegetated land surface would have allowed loss of much of the tracer to plant interception and subsequent delay in the complete entry of the solute into the underlying soil horizons of interest. Eighty grams of $CaBr_2$ were dissolved in approximately 15 L of water and sprinkled onto the surface of the plot as evenly as possible using a garden bucket and slowly enough that the 4-mm equivalent depth infiltrated. Bromide concentration in the applied solution was about 1000 mg $L^{-1}$. Laboratory measurements of bromide content of the source reagent showed that bromide had been added at an average rate of $15.2 \pm 0.09$ g $m^{-2}$. A rain event began just after bromide had been applied at Transect 1 (Sep 26) and prior to application in plots at Transects 2 and 3 (Sep 27), with most of the 24 mm of rain falling before bromide had been applied at these other two plots. Little vegetation re-grew on the plots for the remainder of the study.

Spring and autumn soil sampling of the bromide plots was intended to compare effect of snowmelt and summer precipitation on water movement patterns. Sampling was restricted to within the plot volume because of the prohibitive costs of obtaining the very much larger sample size needed to adequately characterize plume behaviour outside of the plot. Vegetation uptake of bromide was not measured because there was no way to distinguish whether bromide had resided within the plot or had already moved out of the plot at the time it was absorbed by plant roots. Bromide plots were sampled on four occasions over a 751-day period: May 21/98, October 13/98, June 25/99 and October 18/99. In the remainder of this paper these are referred to as Spring/98, Fall/98, Spring/99 and Fall/99 respectively. Sampling was conducted in increments of 0-10, 10-20 cm and each 20 cm thereafter with a 7-cm diameter soil auger. In spring/98, plots were sampled to a 120-cm depth at two points per plot and samples for each increment were combined. For the three remaining sampling events, plots were sampled at 4 points to a 160-cm depth, but samples from the same increment were not combined. Moisture content was determined on all samples by oven-drying a subsample at 105 $^{\circ}$ C. Samples were air-dried, ground to pass 2 mm and stored at room temperature. Bromide was extracted from soil suspension after shaking 1 hour in 0.005 M $CaCl_2$ solution. Extracts were centrifuged, filtered through GF/F filters, and stored at 5 C. Bromide concentration was determined with a Dionex ion chromatograph with an AS-9 column. Analytical detection limit was 0.1 mg $L^{-1}$. Dilution imposed by the extraction process increased the detection limit to about 1 mg $L^{-1}$ in the original soil. Every 4$^{th}$ sample was extracted in duplicate to ascertain whether analytical replication was needed. The resulting linear regression between 84 pairs of samples had an $r^2$ of 0.997 and a slope of 0.97 ± 0.006, indicating high extraction precision.

Interflow was measured at the lower midslope position at each transect (Figure 1) with troughs designed to quantify subsurface flow rates and to obtain soil water samples for chemical analysis. Atkinson (1978) has reported that troughs and particularly open trenches sometimes employed with troughs can disrupt normal flow patterns; nevertheless, interception troughs are still commonly employed in subsurface flow measurement (Wilson *et al.* 1990; Renzetti *et al.* 1992). Rectangular stainless steel troughs were installed at 3 depths in the profile. The collectors in the Ae horizon intercepted the upper and lower 6 cm of the horizon, while the LFH flow collectors enclosed the entire litter layer (Figure 3). Two troughs, each 1.22 m in width, were placed at each increment and flow from the two was combined. Collectors were inserted horizontally upslope from trenches dug to expose soil horizon boundaries (Figure 3). After tubes were attached to the troughs, trenches were backfilled to minimize disturbance of water movement patterns. Troughs were expected to operate only during periods of rainfall or snowmelt when soil water content approached saturated conditions and soil water pressure reached atmospheric levels (Atkinson 1978). Subsurface flow was measured from June 1998 to October 1999, manually with plastic buckets in 1998, and continuously with automatic tipping bucket flowmeters in 1999. Catchment area for the flow collectors was not determined and nor was I able to distinguish among the sources of water intercepted by the uppermost flow collector, which would consist in theory of both vertical and horizontal components of subsurface flow and overland flow associated with the litter layer.

Soil water content was monitored on the hillslope over the period of the study. Subsamples of soil were taken for moisture determination during bromide plot sampling. Liquid water content was also recorded on a continuous basis with Campbell Scientific CS615 time domain reflectometry probes (TDR) installed horizontally at a 15-cm depth in the lower Ae horizon and at a 40-cm depth in the Bt horizon at the upper-mid slope position of each transect by digging holes to the appropriate depth, laying the probes flat, then backfilling with original material. Measurements were recorded hourly from October 1998 to October 1999 and converted to volumetric water content after probes were calibrated for the Ae and Bt soil horizon materials. Soil water content in the upper 1.0 m of the soil profile was monitored by sampling in increments with a soil auger at each slope position on 7 occasions. Water content of soil samples was converted to a volumetric basis with bulk density measurements and expressed on a depth basis (mm $m^{-1}$) by integrating volumetric water content over the 1.0-m profile depth. Groundwater wells were established at the lower slope position of transects to obtain samples of groundwater for a related soil nutrient study. The approximate depth to the groundwater surface was documented at the time of sample collection. Although soil water content data provided a basis for estimating a water balance for the hillslope, I felt the uncertainties inherent in quantifying either deep groundwater seepage or actual evapotranspiration made such an exercise not worthwhile.

### *Results*

Profiles at all locations of the hillslope were classified as Gray Luvisols based on the Canadian system of soil classification (Agriculture Canada Expert Committee on Soil Survey 1987) or as the US Soil Taxonomy equivalent Cryoboralfs (Soil Conservation Service 1975) and had developed on clay-dominated till parent material. Profile morphology was also very similar at all locations of the study hillslope. The uppermost soil horizon was a porous leaf litter layer (LFH) that ranged in thickness from 7 to 18 cm (mean = 11± 0.9) and graded abruptly into the underlying mineral soil usually with a thin charcoal layer between (Figure 4). The first mineral soil horizon (Ae) varied in thickness from 10 to 23 cm (mean = 16 ± 1.3) and graded abruptly into the underlying Bt horizon. The Bt horizon ranged in thickness from 37 to 56 cm (mean = 45 ± 2.2) and graded gradually into the underlying BC horizon. The BC horizon was a transition layer between the horizon of clay accumulation above and the unweathered soil parent material below. The soil parent material was encountered typically from 160 to 200 cm below surface (mean = 179 ± 3.9; data not shown). Subsoil horizons (Bt, BC) and soil parent material (data not shown) were clay dominated (Figure 5) whereas the Ae horizon had a relatively low clay content (mean = 15 ± 1.1%) and a high silt component (mean = 64 ± 2.1%). There were no evident directional trends in either horizon thickness or texture on the study hillslope. Organic carbon content was less than 1% in all horizons but the litter layer (mean = 40.4±2.1%). The Ae horizon had a platy structure, where soil aggregates were horizontally oriented, whereas the underlying Bt and BC horizons had a blocky structure with vertical and horizontal cleavage between soil aggregates. Plant roots were located almost exclusively within the litter layer and the Ae horizon, with only a few roots extending into the subsoil.

Bulk density of the Ae horizon measured the year after harvest had a mean value of 1.65 Mg $m^{-3}$, reflecting recent compaction from logging (Whitson *et al.* 2003b). Bulk density measured three years after harvest is shown in Table 1. Mean value ranged from 1.11 to 1.33 Mg $m^{-3}$ in the first and second 10-cm increments respectively. Average bulk density of the litter layer was 0.15 Mg $m^{-3}$, almost an order of magnitude less than values in the mineral soil. The lower Bt and upper BC horizons were the location of highest bulk density, reaching a maximum of 1.48 Mg $m^{-3}$ at the 80-100 cm depth increment. Porosity as inferred from bulk density measurements and assumed particle density is also portrayed in Table 1. Porosity of the Ae horizon ranged from 50% to 58%, whereas that of the Bt horizon ranged from 45 to 47%.

Saturated hydraulic conductivity ($K_{sat}$) found on the hillslope and from similar soil profiles elsewhere the year after logging are summarized from Whitson *et al.* (2003b). Ranked in order of decreasing median $K_{sat}$ were: Bt vertical ($3.6 \times 10^{-6}$ m $s^{-1}$) > Ae lateral ($1.0 \times 10^{-6}$ m $s^{-1}$) > Ae vertical ($5.7 \times 10^{-7}$ m $s^{-1}$). These values reflected the recent logging activity and particularly the effect of harvest-induced soil compaction. Measurements of steady-state infiltration rate with double ring infiltrometers three years after harvest found that forested sites exceeded logged sites ($5.3 \times 10^{-6}$ m $s^{-1}$ *versus* $2.2 \times 10^{-6}$ m $s^{-1}$) (Whitson *et al.* 2003b).

Precipitation patterns tended towards extremes during and prior to the study (Table 2). While 1997 had above average rainfall, precipitation the next two years was approximately 50% of the average. The largest precipitation event of the study period occurred with snowmelt in 1999, when 87 mm of meltwater infiltrated the soil.

Despite relatively low precipitation, bromide recovery declined with time at all three plots (Figure 6). Declines were particularly dramatic between Fall/97 and Spring/98. Mean bromide recovery ranged from 54% of original at Transect 1 to 29% at Transect 3. In Fall/98, mean recovery ranged from 35±7 % at Transect 1 to just 11±7 % at transect 3. By Fall/99, mean recovery varied from 28±7 % at Transect 1 to just 4±2 % of original application at Transect 3. The amount of bromide remaining in plots appeared to reach an asymptotic value by 1999. A repeated measures ANOVA and Tukey HSD test showed that more bromide was recovered at transect 1 than from transects 2 and 3 in both Spring/99 ($P<0.05$) and Fall/99 ($P=0.05$).

Most bromide resided in upper soil horizons, with concentrations that tailed off to background usually near 80-cm depth (Figure 7). By Fall/99, bromide concentrations were approaching background at all depths at Transects 2 and 3, but not at 1. A bulge in bromide concentration that developed in the 10-40 cm zone at Transect 1 by Spring/98 persisted to later sampling dates. This peak in bromide concentration stabilized in the 20-80 cm zone, a depth corresponding to the clay-dominated subsoil horizons.

Some bromide probably leached beyond the depth of sampling. Detectable concentrations of bromide were found in the lowermost sampled increment at Transects 1 and 2 in Spring/98, and at one of four sample points within each plot for these two transects during the remaining sample dates. Bromide recovery was likely slightly underestimated on the first three sampling dates because the LFH horizon was not sampled. Bromide movement was assumed to result from convection. I lacked data on the hydraulic gradient or the diffusion characteristics of the soil horizons to estimate the 3-dimensional dispersion of the tracer around the plots.

Subsurface flow during the non-snowmelt period was spotty and infrequent in 1998 (Figure 8b-c), but was nonetheless linked to the larger rainfall events (Figure 8a). Only flow for the lower Ae horizon at transects 1 and 3 during 1998 is shown, as the remaining collectors produced negligible flow in 1998. None produced flow in 1999 except for during the snowmelt period (Whitson *et al.* 2003a). Flows from the LFH flow collectors were largest but are not reported because they appear to have underestimated the amount of actual flow interception. The very low volumes obtained from the LFH flow collectors are partially due to instrument error, as I believe that LFH collectors were not measuring all possible flow. For instance, based on measured snowmelt water equivalent and changes in moisture content of the LFH horizon between fall and spring, 10.6 L of snowmelt should have been intercepted at each collector and approximately 2 L of snowmelt should have been stored in the LFH. The other 8.6 L would have been intercepted, and was not, probably because the radius of the exit tube (< 1 cm) was too narrow to accommodate the flow rate from this horizon. Measurements of LFH horizon flow are therefore more useful for illustrating timing rather than magnitude of flow associated with this horizon. Runoff from the Driftwood catchment also reflected weather conditions of the study. Environment Canada (Environment Canada 2002) indicated that annual runoff from the Driftwood river catchment was 22 and 11 mm in 1998 and 1999, down markedly from the long-term average of 119 mm.

Soil wetness during the study period varied considerably in response to precipitation and season. Temporal trends in soil water content at the bromide plots were similar among the 3 plots (Figure 9a-c). Bromide plots were generally wettest during Fall/97 and Spring/98, driest in Fall/98 and Fall/99 and intermediate in Spring/99. Soil water content tended to fluctuate more in upper increments than in deeper ones. The TDR measurements show that soil water content increased sharply in response to snowmelt inputs (Figure 10a-c), probably approaching saturation in the Bt but not the Ae horizon. The subsequent downward trend in moisture content of the Ae horizon was interrupted by several summer rainfall events, whereas the moisture content of the underlying Bt horizon trended downward in spite of these rains. The three locations of TDR measurement trended identically over time in their response to precipitation and evapotranspiration. When profile water storage (mm $m^{-1}$) was compared among locations on the slope, temporal trends are evident during the second year (Figure 11), and there are no trends with location on the slope. Soil water content remained nearly constant in 1998 while in 1999 it decreased, reflecting a second year of drought conditions and the gradual exhaustion of soil moisture reserves by the growing aspen vegetation. Ground water reached a maximum height of approximately 4 metres below surface at the lower slope position (-9 m contour) during the spring of 1999. Groundwater probably was fed by the adjacent lake and likely sloped to the west, flowing under the hillslope into the adjacent watershed. Given its relative depth then, groundwater likely had little influence on hydraulic properties of hillslope soil profiles.

***Discussion***

The physical properties of these Gray Luvisolic soil profiles suggested that interflow was likely to have occurred at this site. Higher clay content and bulk density in the Bt than the Ae horizon means a reduction in $K_{sat}$ and in the downward flux of water at that part of the profile. However, I found that hydraulic conductivity of the Ae and Bt horizon were essentially equal at this site despite pronounced differences in texture and structure. Therefore, Bt horizons at this recently disturbed site would not have restricted downward flux and hence interflow was not observed to any great extent at the Ae/Bt horizon interface. My inability to detect these expected trends in $K_{sat}$ may have several explanations. The soil compaction that occurred during forest harvest and that is evident in the higher bulk densities of Ae horizons observed in logged areas (Whitson *et al.* 2003b) would have reduced the contribution of macropores to hydraulic conductivity. Even planar soil voids associated with soil structural development contribute to hydraulic conductivity

(Schoeneberger and Amoozegar 1990). Coen and Wang (1989) concluded that $K_{sat}$ reached a minimum in the BC horizon of Gray Luvisolic profiles because of a reduction in the strength of soil structure in this part of the profile. Logging then, by reducing $K_{sat}$ of the Ae horizon, may have reduced the probability of interflow occurring.

Bromide loss was much greater than expected, given my findings of minimal $K_{sat}$ differences between Ae and Bt horizons and the low saturated flow that occurred during the relatively dry weather conditions. The greatest decrease in bromide recovery occurred over the first winter when soils were moist. A combination of moist soil conditions and just 33 mm snow water equivalent in the snowpack resulted in the disappearance of about half of the bromide applied. The 87 mm snow water equivalent contributed by the second winter's melt into relatively dry soil had less visible impact on the amount of bromide in the plots. These findings may simply reflect greater potential for lateral water movement and thus solute flux when soils are wetter.

Although little bromide appears to have reached lower soil increments, the loss of bromide cannot only be explained by the process of lateral, downslope water movement. A lack of resources precluded sampling both within and outside the plots to confirm the expected plume migration. Loss of bromide to plant uptake was minimized by disturbing the leaf litter horizon during application and destroying all plant roots in the upper mineral soil. Vegetation (grasses, forbs, shrubs, and trees) did not re-establish on plots until late in 1999. Quantification of bromide in the tissue of plants adjacent to plots would not have been able to distinguish between that bromide absorbed by plant roots from within the plot and that bromide that had been absorbed from outside the plot volume. The largest decline in bromide mass occurred between initial application in Fall/97 (September 26-27) and the Spring/98 sampling (May 21), when leaves were absent, and thus transpiration was minimal. Therefore it is unlikely during this first overwinter period that much was lost to plant uptake. Later losses of bromide may have involved more loss to vegetation, as movement of water and solutes in unsaturated conditions was likely very small due to both small lateral hydraulic gradient and hydraulic conductivity. Other evidence suggests that lateral flow in the Ae was probably reduced by harvest and therefore, flow patterns on this logged site probably became more vertical, with more potential percolation.

The extraordinary amount of bromide transport that was observed under unsaturated soil conditions suggests that some preferential flow may have occurred. The gradual loss of bromide from within the plot volume would suggest that some bromide was located in smaller soil pores that were less vulnerable to convective transport (Steenhuis and Muck 1988; Bache 1990). The behaviour of the plume at transect 1 illustrated how bromide trapped in the abundant smaller soil pore network of the clay-dominated Bt horizon was largely immobile. Precipitation at the time of application probably drove the plume downward into the Bt horizon before bromide could disperse into the smaller pore network of the Ae horizon. Loss of bromide from the Ae horizon with larger average pore size is consistent with the greater mobility of solutes associated with larger pore sizes. It is likely that the bromide movement observed during these conditions represents local soil water redistribution rather than subsurface flow that would contribute to catchment runoff. The infrequent small flows observed by subsurface flow collectors may themselves be examples of local soil water redistribution via preferential flow channels following infiltration.

Soil water content measurements suggested that most bromide transport and subsurface flow occurred under unsaturated conditions. Temporal trends in soil moisture content over both long and short time intervals were consistent within and among different slope positions on the hillslope. This suggests that the water movement patterns inferred from the bromide transport study were likely valid for other slope positions on the hillslope. A bulge in soil water content in the Bt horizon observed in Fall/97, Spring/98 and to some extent Spring/99 at the bromide plots combined with the shallow depth of penetration of the bromide plume suggests that water percolates slowly into these clay-dominated subsoils. It is difficult to apply the term interflow to what was observed to occur on this slope, as the interflow process is usually assumed to occur under saturated conditions.

Quantifying runoff from the hillslope is difficult since the collection area for subsurface flow was unknown. My results indicate that interflow was of very limited importance during 1998-1999 due to low rainfall and relatively low soil water content. Overland flow, if it occurred, was minimal. For the Driftwood river catchment in general, flow in 1998 and 1999 was 19 and 9 % respectively of the long-term mean. Given that Luvisolic soils are one of the dominant soils in regional stream catchments, an increase in annual precipitation would be expected to increase the hillslope contribution to catchment runoff.

### *Conclusion*

Although physical properties of Gray Luvisols favour interflow, changes in hydraulic characteristics of the Ae horizon in terms of higher bulk density and probable reduced vertical hydraulic conductivity affected the occurrence of interflow at this site. Low precipitation and dry soil conditions also influenced the extent of interflow observed. The downward movement of the bromide plume and the increased soil moisture content in the profile confirm that percolation allowed these soils to capture and store water in the soil profile.

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Table 1: Bulk density and porosity of soils of hillslope and adjacent study area

| Horizon or depth increment in cm | | Bulk Density ($Mg\ m^{-3}$) | | Sample Size | | Porosity ($cm^3\ cm^{-3}$) |
|---|---|---|---|---|---|---|
| | | Mean | *CV* | *N* | Pedons | |
| — | LFH | 0.15 | 0.15 | 36 | 9 | 0.85 |
| 0-10 | Ae | 1.11 | 0.14 | 36 | 9 | 0.58 |
| 10-20 | | 1.33 | 0.08 | 38 | 10 | 0.50 |
| 20-40 | Bt | 1.39 | 0.11 | 24 | 6 | 0.47 |
| 40-60 | | 1.47 | 0.07 | 24 | 6 | 0.45 |
| 60-80 | | 1.46 | 0.07 | 24 | 6 | 0.45 |
| 80-100 | | 1.48 | 0.08 | 24 | 6 | 0.44 |
| 100-120 | BC | 1.42 | 0.06 | 24 | 6 | 0.46 |
| 120-140 | | 1.42 | 0.08 | 24 | 6 | 0.47 |
| 140-160 | | 1.40 | 0.10 | 24 | 6 | 0.47 |

Table 2: Cumulative precipitation (mm) for study hillslope

| Period | 1997 | 1998 | 1999 | Long-term mean[1] |
|---|---|---|---|---|
| May 1 - October 31 | 470 | 168 | 179 | 375 |
| winter (snow water equivalent) | — | 36 | 87 | 150 |

[1] Canadian Climate Normals, 1970-1990, Slave Lake, Alberta. (Environment Canada 2000)

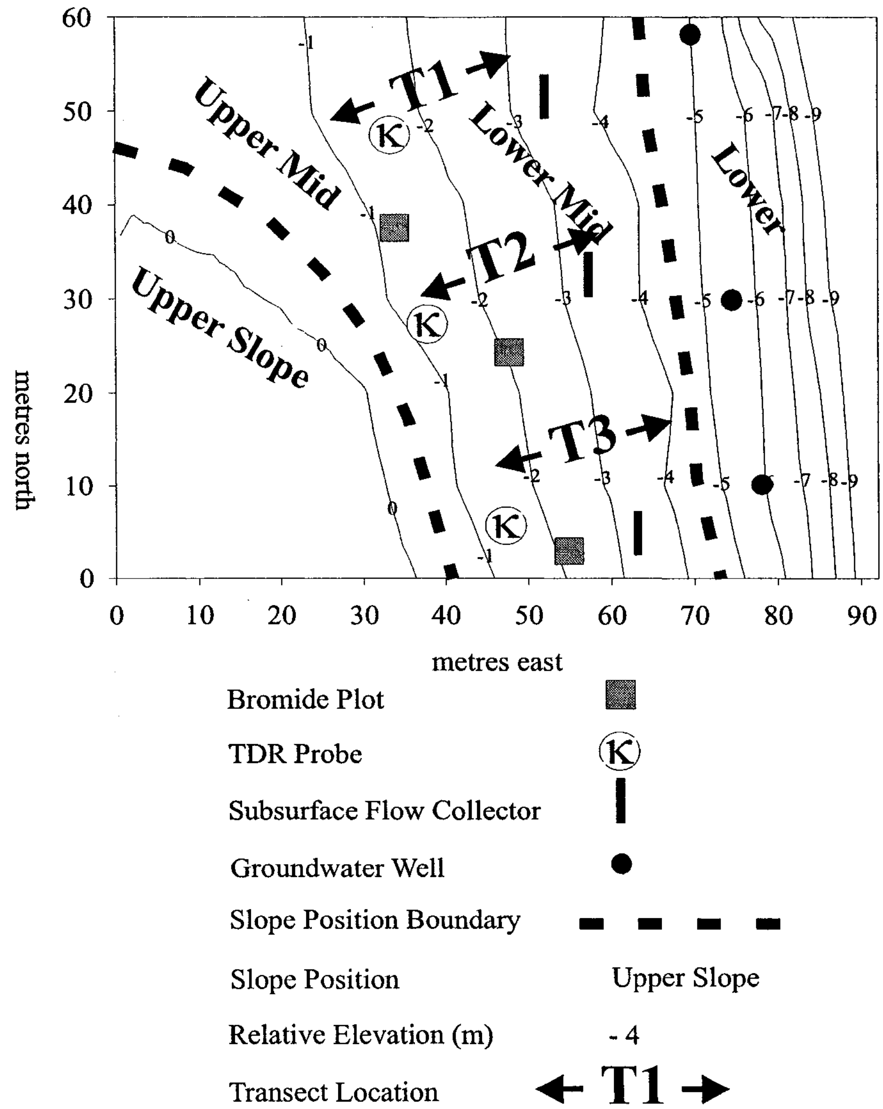


Figure 1: Hillslope diagram showing measurement locations, contours, transects and slope positions

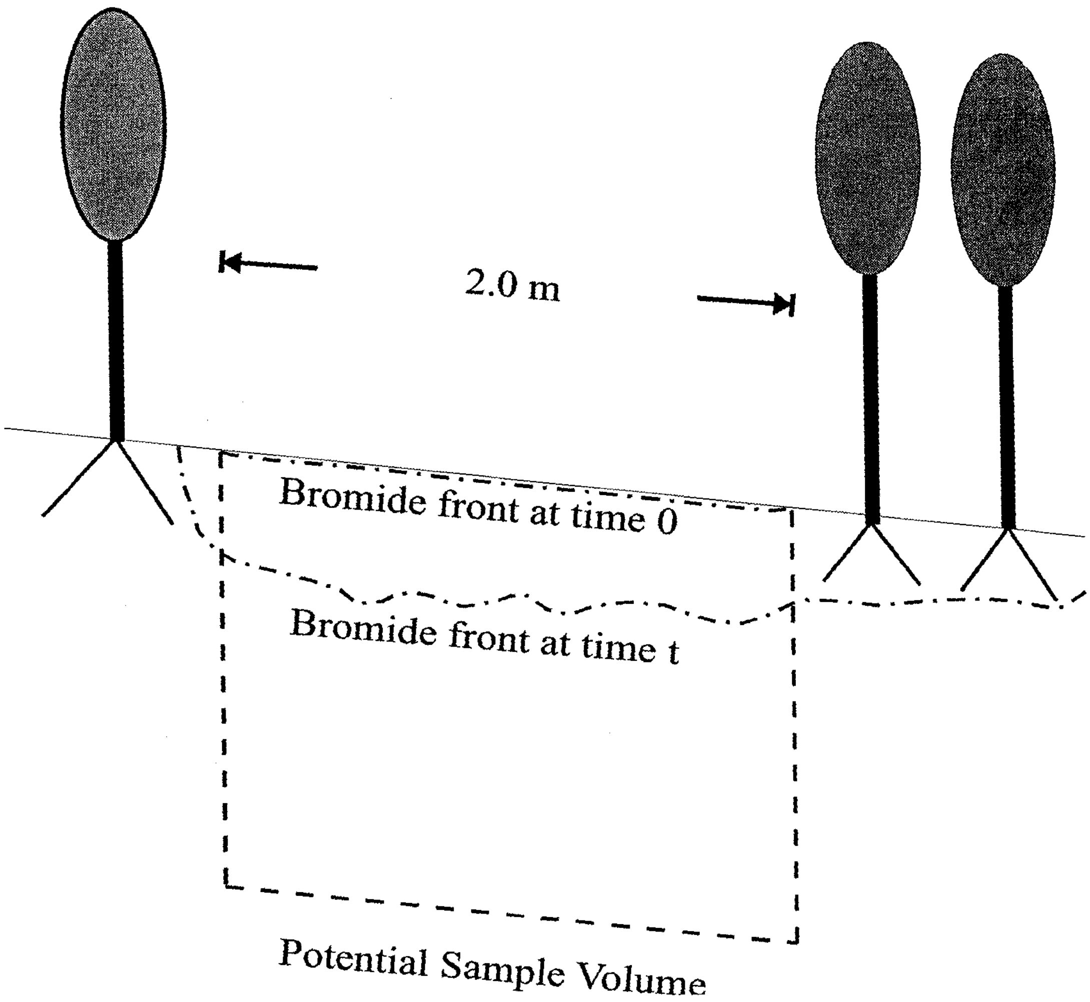


Figure 2: Schematic of 2-dimensional cross-section of bromide plots showing potential plot sampling volume, conceptual position of bromide plume front at beginning of study (t=0) and at some point in time later (t=n)

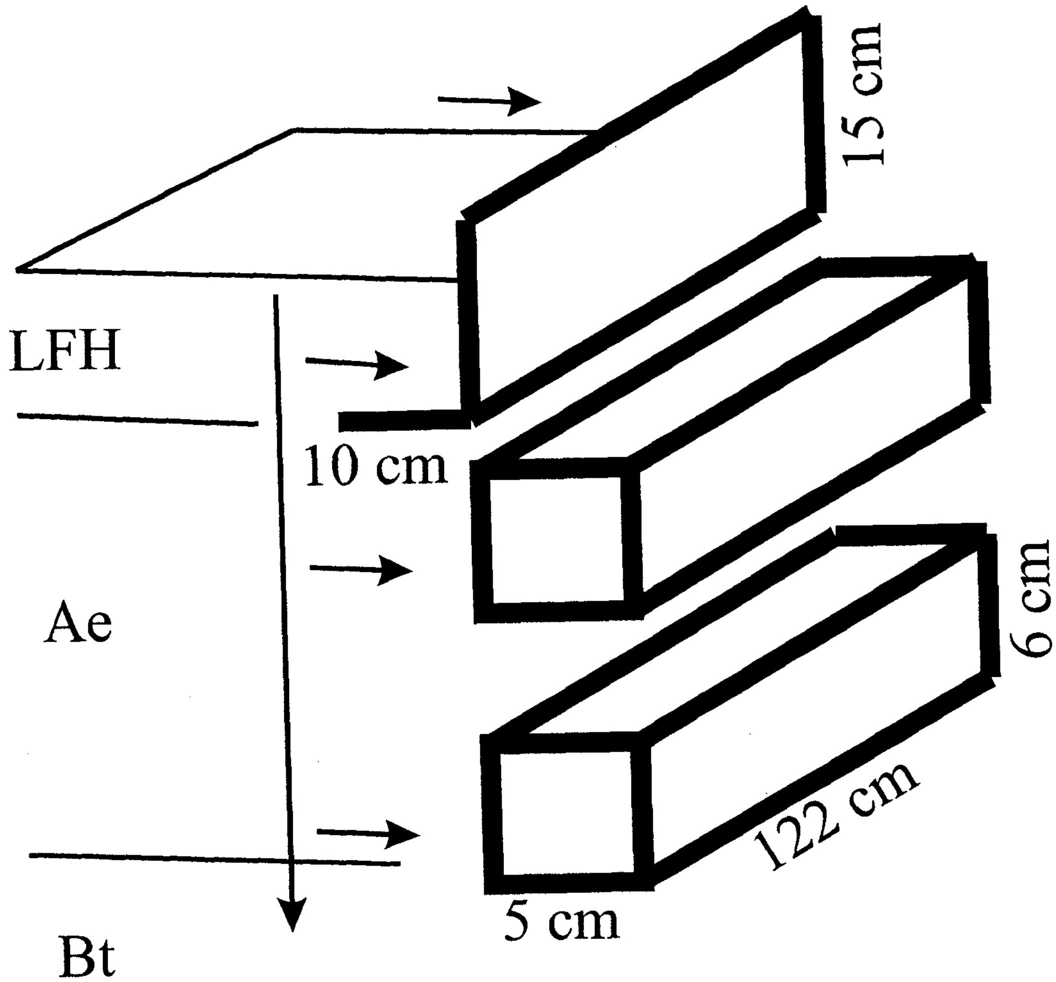


Figure 3: Soil profile diagram showing horizons, potential flowpaths (→) and flow collectors

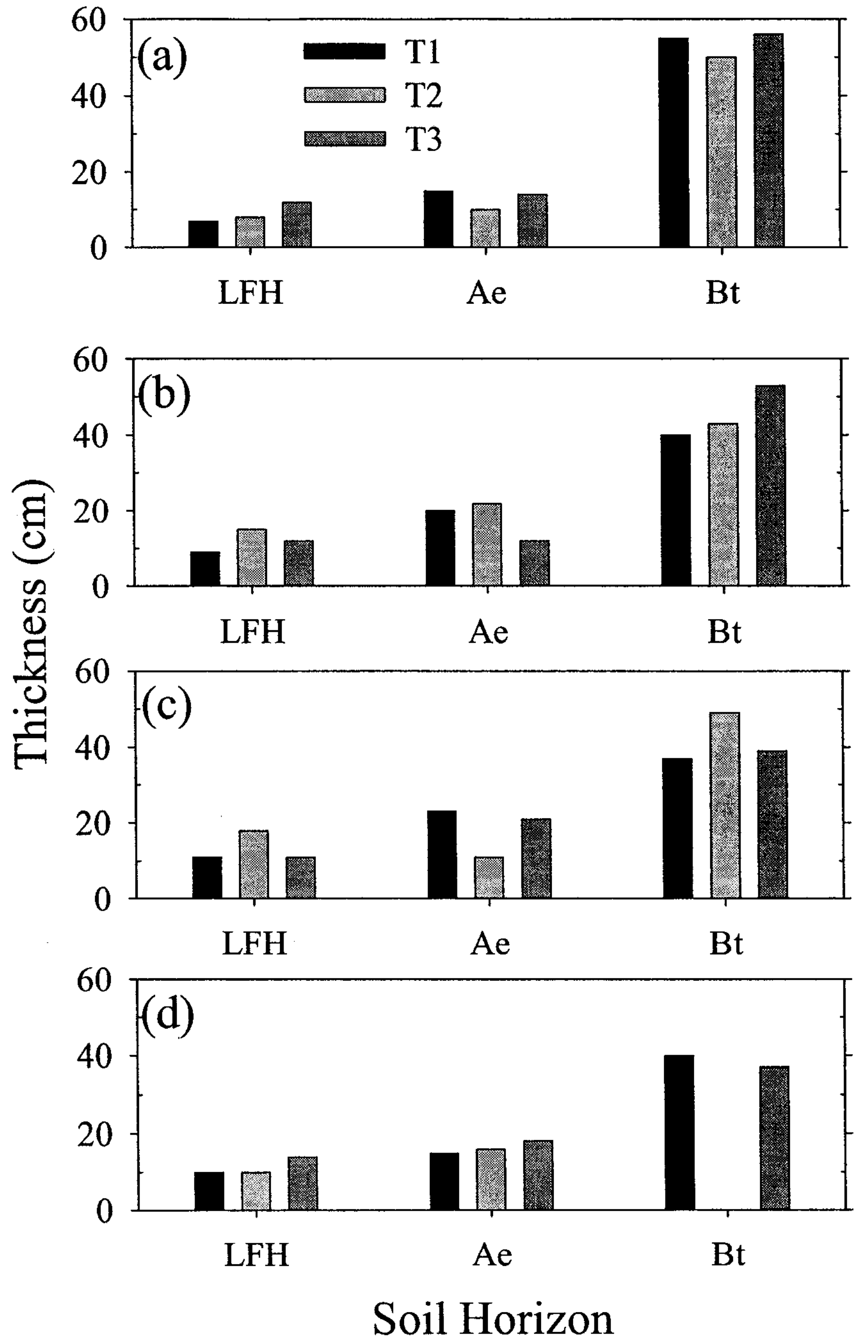


Figure 4: Soil horizon thickness by location on hillslope a) upper slope b) upper midslope c) lower midslope d) lower slope position

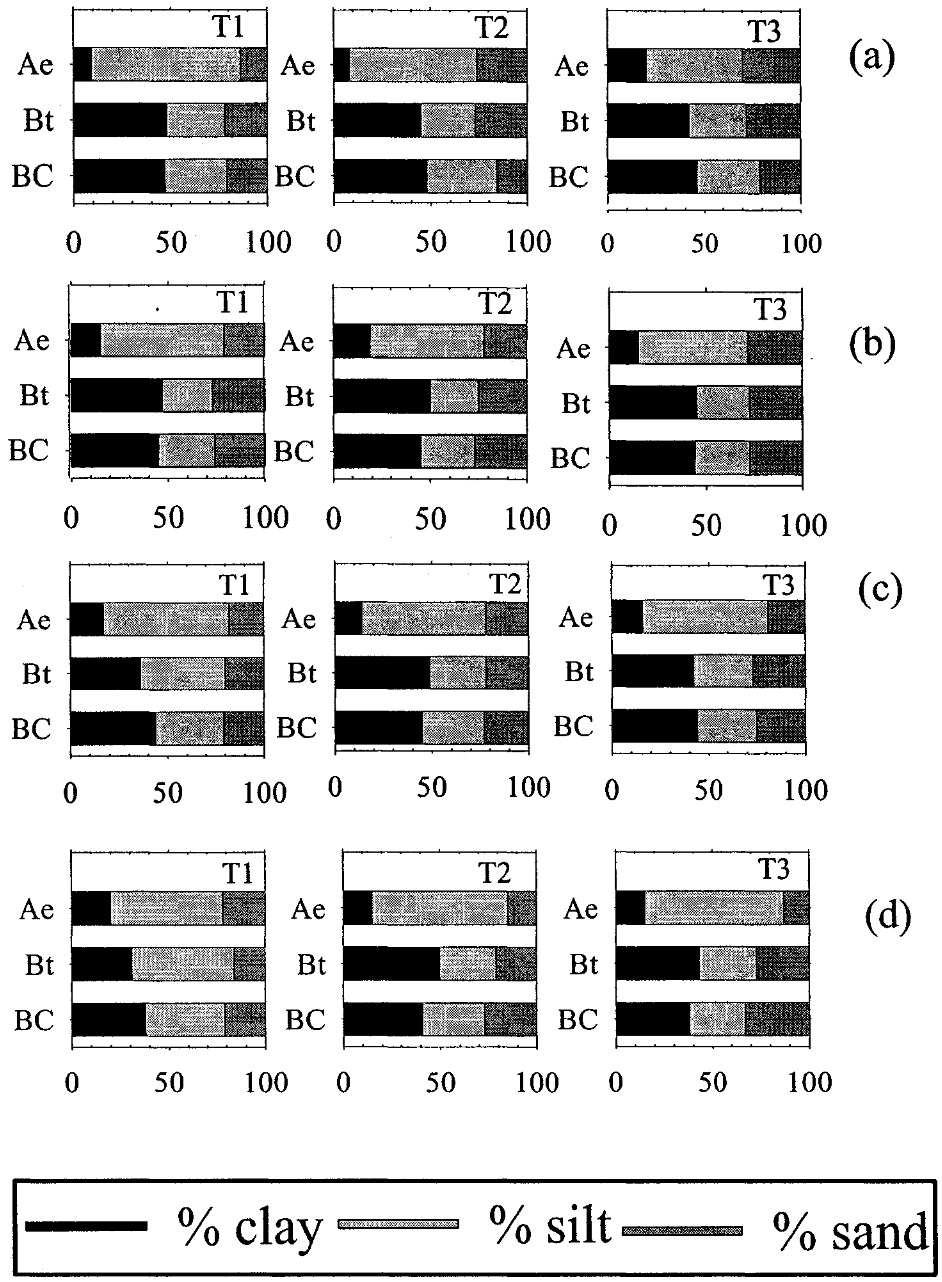


Figure 5: Soil particle size by horizon and locations: (% by weight) a) upper slope b) upper midslope c) lower midslope d) lower slope position

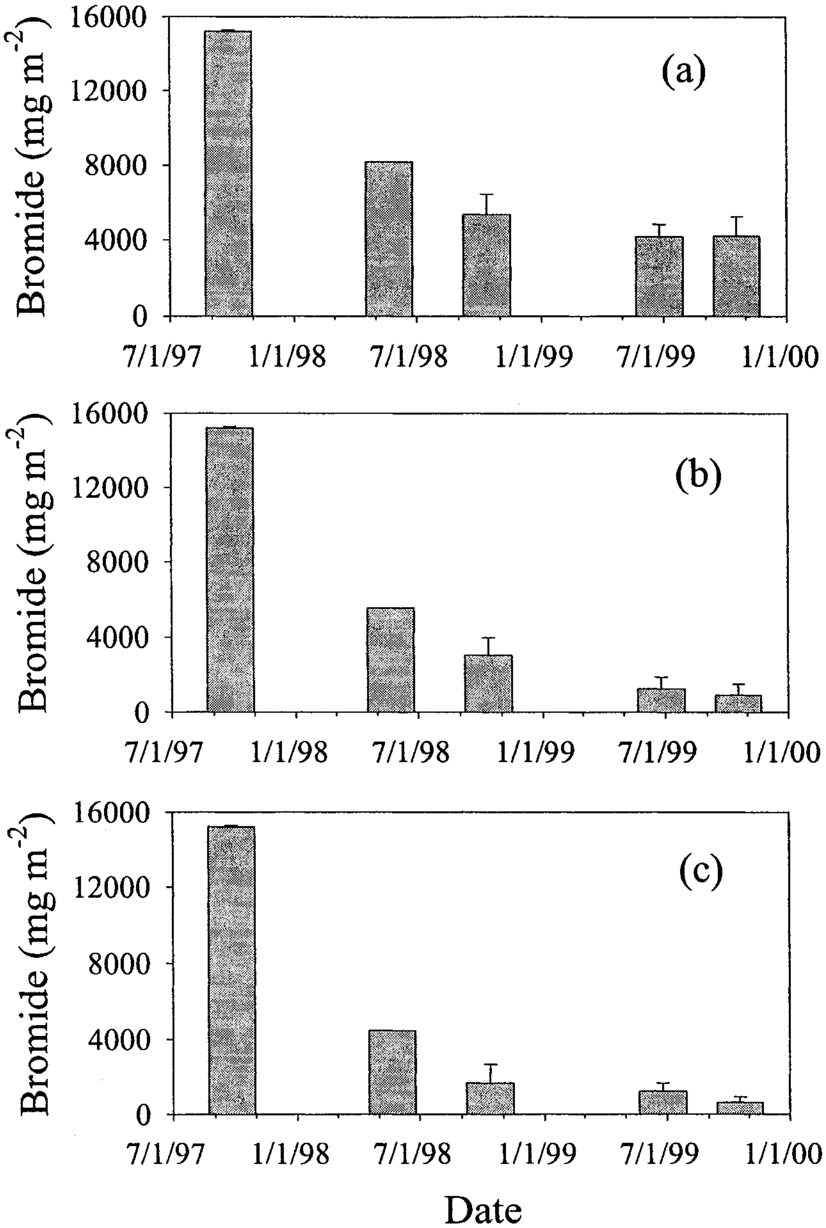


Figure 6: Bromide recovery over time for plots at a) transect 1, b) transect 2, and c) transect 3

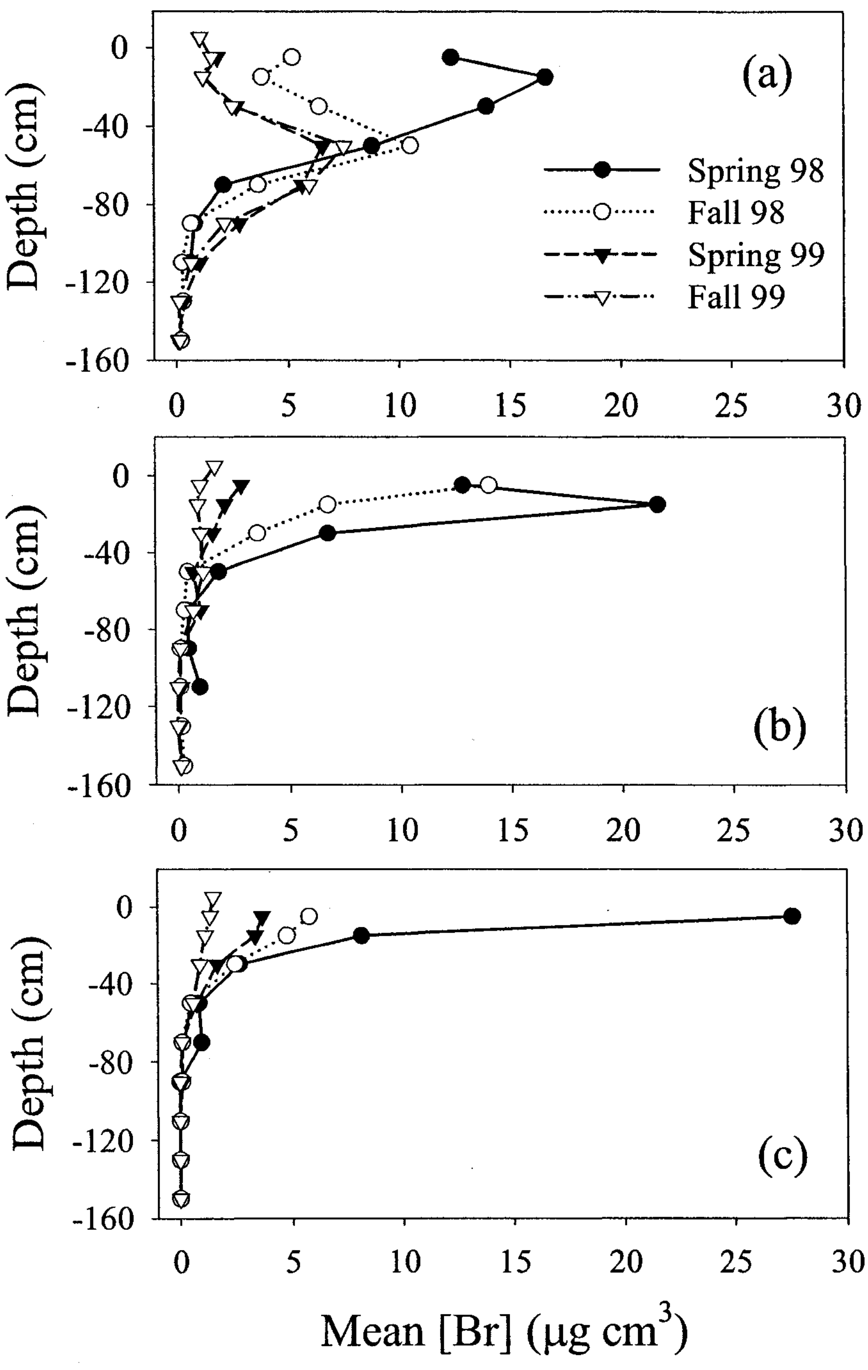


Figure 7: Bromide distribution on a mass/bulk volume basis at a) transect 1, b) transect 2, and c) transect 3

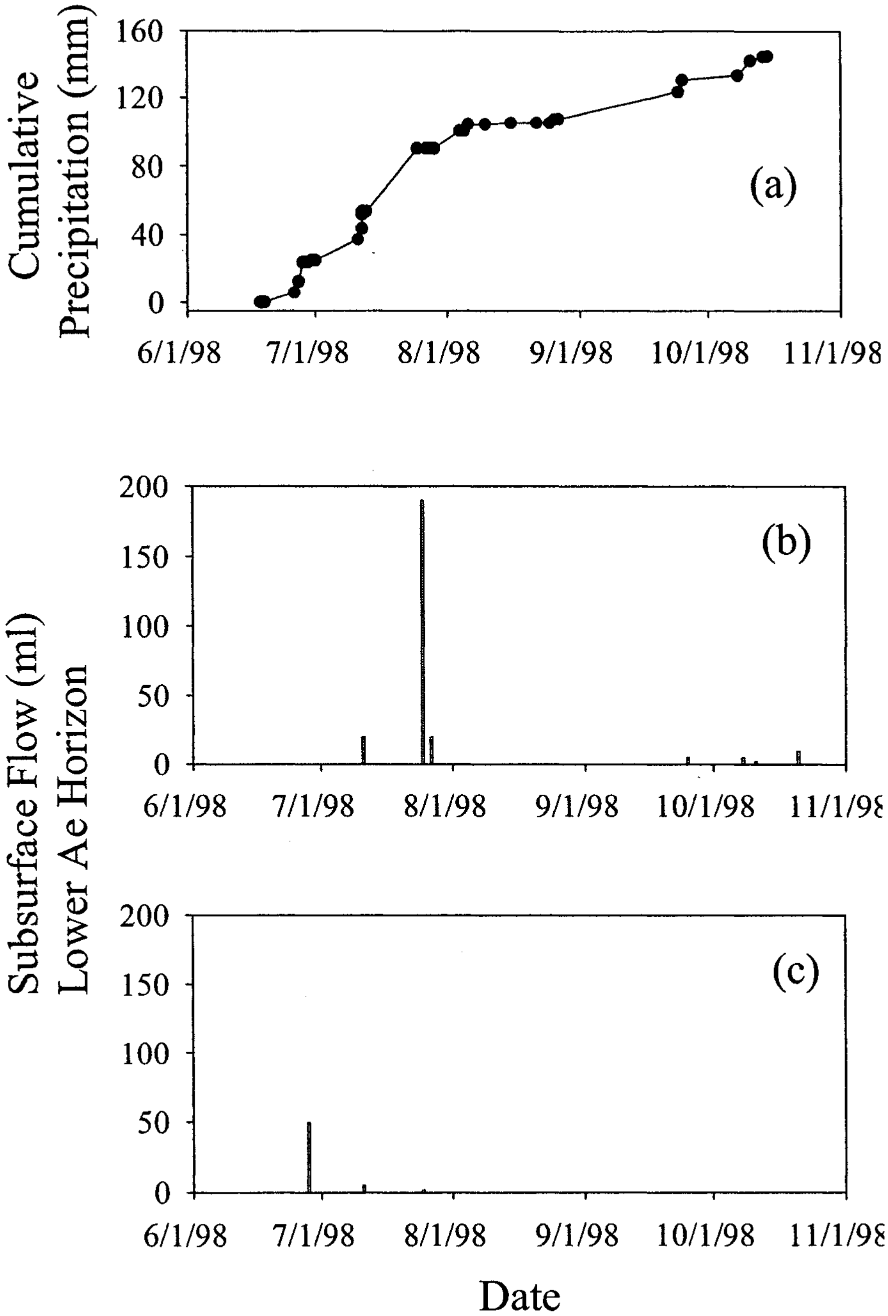


Figure 8: 1998 precipitation and subsurface flow: a) cumulative precipitation at study site, b) subsurface flow in lower Ae horizon at transect 1, c) subsurface flow in lower Ae horizon at transect 3

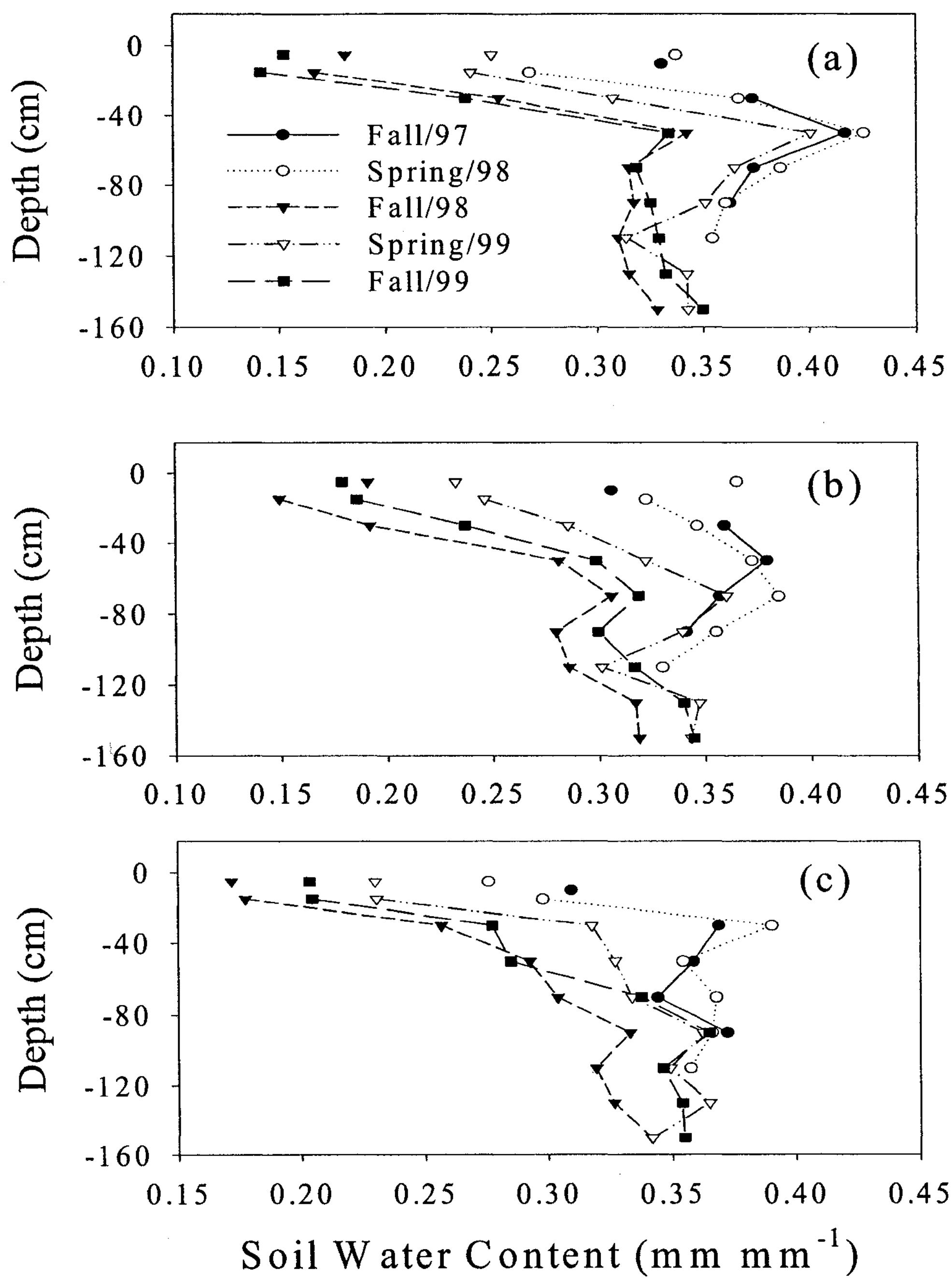


Figure 9: Soil water content at bromide plots over time at a) transect 1, b) transect 2, and c) transect 3

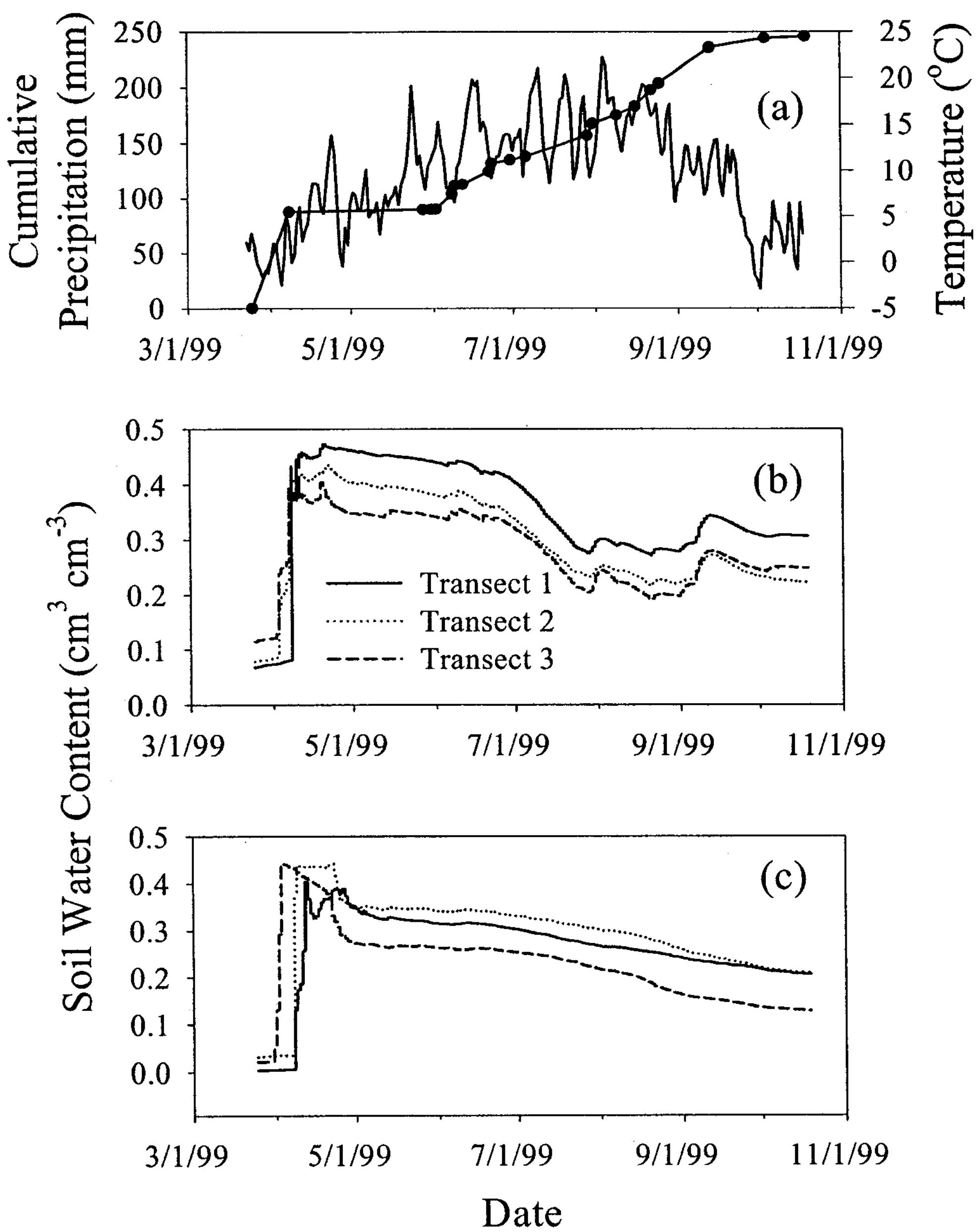


Figure 10: 1999 precipitation, temperature and soil water content relationships: a) cumulative precipitation and mean daily air temperature, b) Soil water content by TDR in Ae horizon, and c) soil water content by TDR in Bt horizon

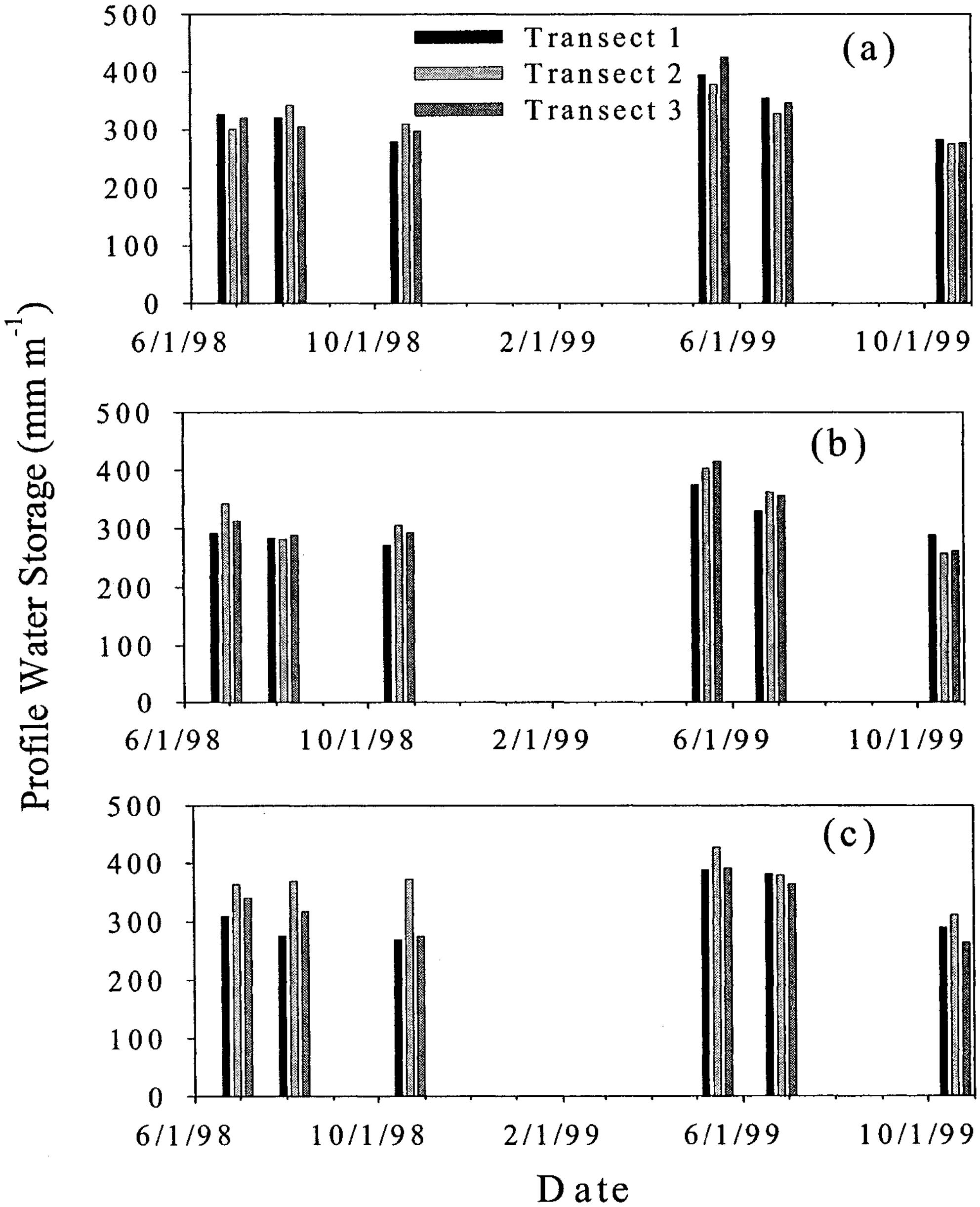


Figure 11: Soil profile water storage by slope position a) upper slope, b) upper midslope and c) lower midslope

## Chapter 3

## Hydraulic properties of Orthic Gray Luvisolic soils and impact of winter logging[1]

### *Introduction*

Forest removal will affect near-surface hydrologic processes on the Boreal Plain. Under forested conditions, infiltration capacities are expected to exceed rainfall or melt intensity such that overland flow will not occur except in areas prone to saturation. Wildfire or timber harvesting, and associated disturbance such as forest-floor scarification or road-building, can expose fine-textured mineral soil horizons to compaction and raindrop impact, ultimately reducing infiltration capacities (Tsukamoto 1975; Cullen *et al.* 1991; Ballard 2000). The longevity of the impact will likely depend on the fire intensity, type of harvest, or the degree of post-harvest scarification. Infiltration-excess overland flow is therefore more likely to occur after these disturbances, particularly with those practices that most reduce infiltration capacity. The profile characteristics of upland forest soils on the Boreal Plain suggest that interflow will be as common as it is elsewhere in North American forest systems (Hewlett and Hibbert 1965).

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[1] A version of this chapter has been accepted for publication as I. R. Whitson, D.S. Chanasyk, and E.E. Prepas. 2003. Hydraulic properties of Orthic Gray Luvisolic soils and impact of winter logging. Journal of Environmental Engineering and Science (in press).

The generation of runoff by overland flow or interflow depends on precipitation patterns however. High intensity storms may exceed infiltration capacities of the surface and thereby create overland flow. Interflow on the other hand requires enough precipitation to create a perched water table in the soil profile. Most rainfall in the Plains region of western Canada results from either convective or frontal storm systems (McKay 1970). Rainfall in the Boreal Plain Ecozone is greatest in June and July, when about 33% of annual precipitation falls (Environment Canada 2000). These months also tend to experience the highest rainfall intensity, with maximum 24-hour rainfall commonly ranging from 50-90 mm (Environment Canada 2000). On an annual basis though, the Boreal Plain Ecozone is characterized by relatively low precipitation and evapotranspiration because of the cooler northern climate. Annual precipitation for the Ecozone is approximately 500 mm, of which about 25 to 33% occurs as snow (Environment Canada 2000). In the Alberta portion of the Ecozone, mean annual moisture deficits range from near 0 at higher elevations to over 100 mm elsewhere (Strong and Leggat 1992).

Interflow has been identified as an important part of the runoff process in many forested locations in North America. Associated with this process is a sharp reduction of hydraulic conductivity at some depth in the soil profile (Whipkey 1965). Interflow has been observed at the soil/bedrock interface in shallow-soil locales of the Canadian Shield (Renzetti *et al.* 1992), within the Bt horizons of Ultisolic soils formed on deeper materials of the southeastern United States (Wilson *et al.* 1990), above poorly-structured subsoils derived from weathered pyroclastic materials in the northwestern United States (Harr 1977), and through fine-textured Bt horizons at higher elevations in the semi-arid western United States (Wilcox *et al.* 1997). A study of interflow in the forested, sub-humid Boreal Plain of northern Saskatchewan found that interflow occurred at the contact between sandy glacial-fluvial and finer-textured till materials (Kachanoski and De Jong 1982). The contribution of naturally occurring soil cements to hydraulic properties has received limited attention even though fragic soil horizons have been often associated with reduction in vertical $K_{sat}$ (Mehuys and De Kimpe 1976; Habecker *et al.* 1990) or with changes to the ratio of lateral to vertical $K_{sat}$ in the profile (Dabney and Selim 1987; Day *et al.* 1998). Fragic horizons typically have high bulk density and strong consistence and occur in Spodosols (Buol *et al.* 1980), Ultisols (Miller 1983) and Alfisols (Rust 1983) and also in relatively young soils developed on till materials (Day *et al.* 1998). Reduction in $K_{sat}$ and root volume along with a sharp increase in bulk density (1.40 - 1.78 $Mg{\cdot}m^{-3}$) reported by Whipkey (1965) below the horizon where most lateral flow occurred may have been due to overlooked fragic conditions. Occurrence of these substances along macropore walls may be equally effective in developing anisotropic conditions within these entities and may partially explain the observations of preferential flow in landscapes where these soils dominate (Wilson *et al.* 1990; Peters *et al.* 1995).

Interflow is likely to be important in upland soils on the Boreal Plain because of the hydraulic characteristics of typical soil profiles. Orthic Gray Luvisols dominate much of this Ecozone (Ecological Monitoring and Assessment Network 1996). A typical horizon sequence consists of an organic LFH horizon, an Ae horizon leached of clay and a Bt horizon of clay accumulation (Agriculture Canada Expert Committee on Soil Survey 1987). Bt horizons are likely to restrict downward flow due to higher clay content and bulk density relative to overlying horizons (St. Arnaud and Whiteside 1964), whereas relatively coarse textured Ae horizons have horizontal or platy structural units reflecting ice-lens formation (St. Arnaud and Whiteside 1964), a condition that may impart anisotropy to them. Soil genesis then has produced a surface with high infiltration capacity and a profile with vertical anisotropy.

### *Objectives and Hypotheses*

The objectives of this study were to determine the effect of timber harvesting during winter on the infiltration process and to examine the potential for interflow in fine-textured Orthic Gray Luvisolic soils. I hypothesized that:

- Harvested sites would have lower infiltration capacities than forested sites
- Ae horizons would have higher infiltration capacities than Bt horizons
- Bt horizons would have lower vertical saturated hydraulic conductivity than Ae horizons
- Lateral hydraulic conductivity of the Ae horizon would exceed vertical.

I measured infiltration with double ring infiltrometers on forested and recently harvested areas. Saturated hydraulic conductivity of soil horizons from several logged pedons was measured with undisturbed Uhland cores and a falling head permeameter.

### *Methods*

*Site Conditions*

Research was conducted at two locations about 120 km apart in the Gray Luvisolic-dominated Boreal Mixedwood Ecoregion of northern Alberta, Canada (Strong and Leggat 1992). Mean air temperature for the period 1952 to 1990 at Athabasca, Alberta, midway between the two locations was –15.7 and 16.3°C in January and July, respectively, and mean annual precipitation was 501 mm, of which 27% occurred as snow (Environment Canada 2000). The maximum-recorded 24-hr precipitation of 84.8 mm was recorded in July (Environment Canada 2000). Aspen (*Populus tremuloides*) dominated the forest canopy prior to harvest at the Lac La Biche (LLB) site (55.2° N, 111.8° W). The forest canopy prior to harvest at the Pelican Hills (PHS) site (55.4° N, 113.7° W) was dominated by both aspen and white spruce (*Picea glauca*). Depth to sedimentary bedrock at the locations was unknown but for the region ranges from less than 15 m to greater than 150 m (Pawlowicz and Fenton 1995). Great depth of unconsolidated deposits meant that percolation was a potential pathway for water movement. Orthic Gray Luvisolic profiles chosen for this study belonged to the Grandin and Athabasca series (Knapik and Brierley 1993). Parent materials of these respective series consist of fine and moderately fine textured till. Upland areas were harvested in January, 1997, with all coniferous and deciduous trees removed from blocks that ranged in size from 10 to 40 ha. Observations at the time of harvest suggest that soils were frozen but the depth of frost was unknown. Snow depth was approximately 40 cm. Logging was carried out by a combination of cut-to-length and whole-tree-length techniques. With the former, trees were processed at the stump with a Timber Jack 1270 and wood was hauled to a logging road with a Timber Jack 1210

Forwarder. With whole-tree-length harvest, a Timber Jack 618 Feller-Buncher cut the trees and a grapple skidder moved the logs to roadside for de-limbing and subsequent loading. Harvested areas received no further post-harvest silvicultural treatment other than the planting of white spruce seedlings.

*Measurement*

Infiltration was measured at the PHS site the third summer after harvest with double ring infiltrometers (Gray *et al.* 1970). Diameter of the inner and outer ring was 310 and 520 mm, respectively, and the double rings served to ensure one-dimensional flow. Water obtained from a nearby lake was added to both the inner and outer rings, and the rate of fall of water in the inner ring was measured while the water level in the outer ring was kept filled. Infiltration measurements were made at 9 sites, 4 under logged conditions and 5 under forest within a 2-$km^2$ area. Measurements were made at up to 12 pedons at each site. Prior to inserting the rings, the soil was excavated to expose either the Ae or Bt horizon surface. Rings were driven an average of 95 mm into the soil for Ae horizons and 76 mm for Bt horizons. Infiltration was measured at times (t) of 0.5, 1, 1.5, 2, 3, 4, 5, 7, 10, 15 and 30 minutes and each 30 minutes thereafter until 180 minutes or longer. Infiltration measurements were conducted only at the PHS site because the Luvisolic soils at both locations of PHS and LLB belonged to soil series that were very similar in terms of soil texture (moderately fine *versus* fine).

The empirical Kostiakov equation (Jury *et al.* 1991) was applied to infiltration data because the plot of log I vs. log t was linear (Figure 12b and 13b):

[1] $I = at^m$ or $\log I = \log a + m^* \log t$

where the intercept a is the cumulative infiltration at t = 1 minute and m is the slope of the relationship, or in effect, an index of the infiltration rate. The two derived parameters for intercept (a) and slope (m) of the log-I vs. log t relationship can be compared to test for differences in infiltration characteristics between treatments, as was done for rangeland sites in Alberta (Naeth *et al.* 1991).

Uhland soil cores (76-mm diameter and length) were collected for hydraulic conductivity and bulk density measurement at both PHS and LLB locations. Two soil pits (~1.5 $m^3$) at the LLB site and four at the PHS site were excavated in the summer of 1997. Six cores each were obtained from vertical and horizontal orientations of the Ae horizon, while six cores were obtained from vertical orientation of the Bt horizon. Cores from the Bt horizons of 3 PHS soil pits could not be obtained due to the combination of fine textures and moist soil conditions. Saturated hydraulic conductivity ($K_{sat}$) was determined in the laboratory with a falling head permeameter after slowly saturating soils from below over a period of several days to remove trapped air (Reynolds 1993). Bulk density was measured by oven drying Uhland cores after completion of $K_{sat}$ measurement. A second set of measurements was obtained in 1999 at 10 soil pits ranging in size up to 2 $m^3$ at the PHS location with Uhland cores. Samples were taken in 100-mm increments from the Ae horizon and 200-mm increments below that to 1600 mm.

In addition to these $K_{sat}$ and infiltration measurements, a single measurement of lateral $K_{sat}$ of an Ae horizon was made at the PHS site. A zero-tension subsurface flow collector (Atkinson 1978) buried at the Ae-Bt horizon interface and used to obtain soil-water for chemical analyses experienced a period of saturated subsurface flow during the 1999 snowmelt. Darcy's law was used to calculate $K_{sat}$ from the subsurface flux rate, cross-sectional area of the flow collector, and slope gradient at the location (Weyman 1973). $K_{sat}$ calculations were based on a flux of 500 $ml \cdot hr^{-1}$ that remained constant for a 19-hour period through a cross-sectional area of 0.1464 $m^2$ under a gradient assumed equal to the sine of the slope angle of 7.4 $^{\circ}$ (Harr 1977).

Soil samples were obtained to determine gravimetric water content at 11 additional pairs of forested-logged sites at PHS by taking increments of soil to a 1000-mm depth and converting these to volumetric water content with measures of bulk density. These samples were obtained for a related study that compared soil moisture content between forested and harvested sites. Particle size was measured by the hydrometer method (Sheldrick and Wang 1993) with samples from a subset of the LLB and PHS pedons used in the study.

*Statistical Analyses*

The relationship between cumulative I and t was summarized within treatments by linear regression to develop a single best-fit line from replicate values of I at every value of t (Zar

1996). Differences in Kostiakov parameters between treatments were tested for statistical significance by an Analysis of Covariance procedure (Zar 1996), with treatment (horizon x disturbance) the predictor variable and time the covariate. If slopes of regression lines to $t = 180$ minutes differed among treatments a post-hoc Tukey test was employed to determine which treatments were different. Y-intercepts of those treatments that had the same slope were then compared by a Tukey test.

$K_{sat}$ values determined by falling head permeameter were first log transformed to allow parametric statistical analysis. Each combination of horizon and core orientation was considered a treatment so that there were 3 treatments: Ae vertical, Ae horizontal, and Bt vertical. Differences in saturated hydraulic conductivity among horizons was determined by a Model 3, two-factor ANOVA, with horizon the fixed factor and soil profile the random factor. $K_{sat}$ values of Luvisolic profiles approximated from the slope of the I *vs.* t relationship at $t \geq 60$ minutes obtained by infiltration measurements were also log transformed and compared by a two-tailed *t*-test.

***Results***

Ae horizons were typically enriched in silt and depleted of clay in relation to the underlying Bt horizons (Tables 3 and 4). These differences are a reflection of downward clay transport (Howitt and Pawluk 1985) and of physical weathering of larger sand particles. The bulk density of Ae horizons sampled in 1997 for $K_{sat}$ analysis (Table 4) was greater than those obtained for the corresponding 0-200 mm depth in 1999 for the infiltration study (Table 5), indicating a reduction in bulk density over the period. In 1997, bulk density of the Ae was on average 1.63 Mg $m^{-3}$. In 1999, the corresponding 0-200 mm depth increment had a bulk density of from 1.22 to 1.33 Mg $m^{-3}$. The big difference over the two years suggests that soils were compacted by harvest and were now slowly returning to background condition.

The Kostiakov equation fit the infiltration data very well, as shown by the high value of the mean $R^2$ coefficient for individual treatments (Table 6) and the linearity of the relationship (Figure 12b and 13b). The Kostiakov slope m differed between horizons regardless of disturbance ($P < 0.005$), with higher values found for forested and logged Bt horizon treatments than for forested and logged Ae horizon treatments (Table 6). Intercepts (a) differed between forested and logged treatments for both Bt and Ae horizons ($P < 0.001$), with lower values found for logged treatments. The difference in intercept value meant that for a given horizon, cumulative infiltration was lower in logged areas than forested at any value of t.

Plots of I *vs.* t followed a typical pattern of initial curvilinear increase to about 30 minutes followed by a period of linear increase to 180 minutes for both Ae (Figure 12a) and Bt surfaces (Figure 13a). The slope of the linear portion of this relationship represents steady-state flow conditions that develop after all soil pores are filled and water moves largely by gravity (Jury *et al.* 1991), and may be considered to approximate $K_{sat}$ (Buttle and House 1997). Under these conditions of downward gradient near unity, the flux may therefore be considered to approximate $K_{sat}$. Log-transformed values of steady-state flux rates calculated for the period from 60 minutes onward were compared between forested and logged Ae horizons (data not shown). Linear relationships were very strong, with a median $R^2$ of 0.996 and 0.994 for forested and logged treatments, respectively. A two-tailed *t*-test found that the geometric mean of steady-state flux of water through the Ae horizon surface was about 2.5 times greater in the forested than the logged treatment ($P < 0.02$) (Table 7).

Soil water content as expressed by saturation percentage at proximate plots was lowest above the 200-mm depth, which corresponds to the Ae horizon, reflecting greater uptake by plant roots near the surface and lower available water holding capacity of the Ae horizon (Figure 14). Water content seemed to reach a maximum at the 500-mm depth, which corresponded to the middle-Bt horizon. Water content on any given sampling period and for a given depth was less on forested sites than logged ones.

During $K_{sat}$ measurement many cores developed piping through small fissures; these were discarded. As a result, measurements were obtained on only 46 of the 91 samples collected. The geometric mean of $K_{sat}$ was unexpectedly highest in the Bt horizon treatment, about 6 times larger than vertical Ae $K_{sat}$ and 3.6 times larger than lateral Ae $K_{sat}$ (Table 7). Treatment differences were not detectable ($P = 0.234$) because of the high standard deviation within each treatment, although interaction between soil pit and horizon was significant ($P$=0.016). $K_{sat}$ determined from subsurface flow collector interception was calculated to be 7.3 x $10^{-6}$ m·s$^{-1}$. This value of $K_{sat}$, although 7.3 times larger than the corresponding value determined by permeameter, may lie within the range of variability normally encountered with this parameter, particularly since the standard deviation for lateral Ae horizon $K_{sat}$ was close to an order of magnitude of the mean (Table 7).

### *Discussion*

#### *Effects of Forest Removal*

The hypothesis that logging would affect infiltration rates was supported by my findings. The greater steady-state infiltration rate in the forested treatment demonstrated that logging still affected hydraulic properties in these soils 3 years post-harvest. Likewise, initial infiltration rate was affected by disturbance, with higher values under forest for both Ae and Bt horizons. These differences in infiltration characteristics between forested and logged sites are likely to have some impact on water movement patterns, at least at the site level. When observed steady-state infiltration rates are expressed in units typically used for precipitation they are 19.1 and 7.9 $mm \cdot hr^{-1}$ for forested and logged treatments respectively. One-hour storms of this range of magnitude have a return period of 5-10 yrs in northern Alberta (McKay 1970). Therefore, infiltration excess overland flow is likely to occur with greater frequency in logged areas where mineral horizons are exposed.

I found limited support for the hypothesis that infiltration capacities would be greater in Ae than Bt horizons in spite of the much finer textures in the latter. The Kostiakov slope parameter was higher for Bt than Ae horizons, indicating that infiltration rates were higher in the Bt horizon treatments than the Ae horizon regardless of disturbance. Differences probably reflect differing structural characteristics between these horizons, with Ae horizons commonly possessing horizontally-oriented structural units, and Bt horizons possessing vertically-oriented structural units.

Differences in bulk density attributable to compaction probably contributed to observed differences in infiltration characteristics between logged and forested treatments. Although these sites were harvested under winter conditions when soils were possibly frozen, and not scarified afterwards, compaction was still evident after 3 yrs in the 0-100 mm increment and possibly in the 200-400 mm increment as well, both increments corresponding to upper Ae and Bt horizons respectively. Bulk density of the 2 upper mineral soil increments measured in 1999 are much lower than the 1997 values for the Ae horizon to which they correspond, probably because of freeze/thaw cycles and other natural amelioration processes.

Differences in infiltration between logged and forested treatments may also have been due to differences in soil moisture conditions. Logged treatments were wetter than forested. All profiles reached a constant wetness at about 500 mm, within the lower Bt. These differences would have affected both the hydraulic gradient and hydraulic conductivity, the latter through soil swelling and bulk density changes. For instance, the vertical hydraulic gradient at $t = 0$ would have been much greater in drier soil than moist. Soil swelling resulting from increased moisture content would have reduced bulk density and increased micro-porosity at the expense of macro-porosity.

Both compaction and increased water content probably reduced soil macro-porosity, the former through compressive force, the latter through expansive forces associated with clay platelet expansion. Macro-porosity reduction would be especially important in the finer-textured Bt horizon. Although vertic behaviour was not observed in either the Grandin or Athabasca soil series, the general preponderance of smectite minerals in Alberta subsoils (Pawluk 1961; Abder-Ruhman 1980; Howitt and Pawluk 1985) means that in particular the fine textured Grandin soil, (> 40% clay content) would undergo large volume adjustment with changes in moisture content. These soils would have lower macro-porosity when wet than dry. Within Vertisolic or vertic suborders of other soil groups steady-state infiltration rate has been shown to depend very much on the initial water content of the profile, with wetter soils having lower steady-state values (Lin *et al.* 1998). Under drier conditions, soil shrinkage would lead to the expansion of inter-aggregate planar voids and higher final infiltration rates. Since infiltration measurements were conducted during a low-rainfall period forested sites in particular were dry and prone to soil shrinkage along ped faces. Hence macro-porosity increase was likely greatest in the drier forested Bt horizons and led to higher rates of water transmission.

The implications to managing forests are that compaction and soil moisture increases associated with logging together reduce macro-porosity and in turn reduce infiltration capacity. Infiltration capacity is likely to be reduced until soil water dynamics approach that of an older forest and porosity returns to normal. I am uncertain how many freeze-thaw and wet-dry cycles this may require, but results from a study on similar soils in northern Saskatchewan found that final infiltration rates had largely converged between forested controls and sites that had been logged and scarified from 4 to 15 yrs earlier (Elliott *et al.* 1998).

*Interflow in Gray Luvisols*

The hypothesis that the Bt horizon would have lower $K_{sat}$ was not supported by my data. The trend towards higher $K_{sat}$ in the Bt than the Ae horizon as determined by permeameter measurement was surprising because clay content was much greater than in the Ae. The high bulk density of Ae horizons relative to the undisturbed state suggests that upper horizons had been compacted by recent forest harvesting and that my measurements of $K_{sat}$ for Ae horizons may not be typical of undisturbed forests. However, even in the undisturbed state there may be less difference between Ae and Bt horizon $K_{sat}$ than would be expected. Coen and Wang (1989) reported that vertical $K_{sat}$ of Ae and upper Bt horizons were similar in three Luvisolic pedons closely related to those of this study and asserted that in Luvisolic profiles downward flow is restricted by BC horizons where soil structure is weaker. In northern Saskatchewan, where interflow occurred at parent material discontinuities, Kachanoski and Dejong (1982) found that $K_{sat}$ of soil cores from weakly-structured C horizons of sandy-clay loam-textured till ranged from $2.8 \times 10^{-9}$ to $2.8 \times 10^{-8}$ $m \cdot s^{-1}$. Weaker soil structure has also been found to reduce $K_{sat}$ in partially weathered saprolite materials (Schoeneberger *et al.* 1995). The higher value of $K_{sat}$ in the Bt horizons relative to the Ae horizons in this study was probably due to the combination of compaction of the Ae and sub-angular blocky soil structure in the Bt. Soil structure may ultimately dominate the other factors of texture or bulk density in determining *in situ* $K_{sat}$ of soil profiles.

My measurements of $K_{sat}$ do not provide support for the existence of interflow in Gray Luvisolic soil profiles despite the similarity of textural and bulk density properties of

Luvisolic soil profiles to soils where interflow occurs extensively. Interflow has been commonly observed in regions of humid climate where soils have structural, textural and bulk density changes in the profile that lead to restrictions in downward flux (Whipkey 1965; Harr 1977; Wilson *et al.* 1990). Given the right amount and timing of precipitation, interflow has been found to occur even in semi-arid climates (Newman *et al.* 1998). There may be several mechanisms that could cause interflow in Orthic Gray Luvisolic soils. Interflow could be a transient phenomenon that occurs while the wetting front crosses the Ae/Bt horizon interface, and may occur most frequently during periods of low intensity rainfall or snowmelt into initially dry profiles. Alternatively, it may be the underlying BC horizon that limits $K_{sat}$ (Coen and Wang 1989), limiting interflow to periods when the profile has been wetted to this depth and saturation has developed upwards into the Ae.

The data on horizontal $K_{sat}$ support the hypothesis that the Ae horizon is anisotropic, that is, hydraulic conductivity at that point is not the same in each principle direction. A ratio of lateral/vertical $K_{sat}$ greater than one means that lateral flow is potentially enhanced by the anisotropy within the horizon. The ratio of lateral/vertical $K_{sat}$ in the Ae horizon ranges from 1.75 if the geometric mean of Ae horizon lateral $K_{sat}$ (Table 7) is used and 12.8 if the single datum from the *in situ* measurement of lateral $K_{sat}$ with the zero-tension flow collector is used. While Gray Luvisols have experienced limited chemical weathering (Pawluk 1961), the fabric of the Ae horizon reflects the platy-structure observed in the field. Within the horizontally oriented platelets, finer material tends to concentrate near the top of each unit (St. Arnaud and Whiteside 1964). Interflow may thus occur in Gray Luvisols because of anisotropy within the Ae horizon, not because of reductions in vertical $K_{sat}$ between the Ae horizon and underlying finer textured subsoil. Interflow has been found to occur within fine-textured horizons in some locales. In the semi-arid western United States, interflow occurred in parts of the profile that were not expected based on measurements of physical properties (Wilcox *et al.* 1997). $K_{sat}$ of the underlying weathered tuff was 3 to 4 orders of magnitude greater than that of the smectite-dominated Bt horizon through which most of the lateral flow occurred. Likewise, flow occurred through finer-textured Bt horizons in the Ultisolic landscapes that Wilson et al. (1990) studied. Bt horizons by virtue of their blocky structure may also be anisotropic, but because vertical interpedal facial area is twice as great as horizontal when vertical and horizontal ped dimensions are equal, these horizons should have a lateral/vertical $K_{sat}$ ratio of less than 1, with greater vertical $K_{sat}$ than horizontal. Hence, interflow is unlikely to be important in this horizon.

### *Conclusions*

Differences in infiltration characteristics caused by logging are still evident three years after winter harvest. These differences probably reflect a reduction in macro-porosity in upper soil horizons due to logging-induced compaction and to increased soil swelling arising from higher moisture content. Some evidence was found to support the occurrence of interflow in Luvisolic soils in the non-frozen post-snowmelt period. Infiltration capacity and $K_{sat}$ of Bt horizons though did not differ as much as expected based on bulk density and textural considerations. Soil structure likely plays an important role in these profiles, allowing faster water movement through the Bt horizon than would be expected based on texture and bulk density considerations alone, and promoting lateral flow within the anisotropic Ae horizon.

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**List of Symbols and Abbreviations**

A intercept

I infiltration

$K_{sat}$ saturated hydraulic conductivity

LLB Lac La Biche

Mg Megagrams

PHS Pelican Hills

SE standard error of the mean

T time

Table 3: Arithmetic mean of particle size distribution of soils from Pelican Hills used for infiltration study

| Treatment | | Particle Size Distribution* Mean % ± 1 *SE* | | | | | |
|---|---|---|---|---|---|---|---|
| | | Ae Horizon | | | Bt Horizon | | |
| | *N* | Sand | Silt | Clay | Sand | Silt | Clay |
| Forested | 5 | 13±1.5 | 71±1.3 | 15±2.2 | 23±2.6 | 36±2.7 | 41±0.4 |
| Logged | 4 | 14±3.8 | 63±6.1 | 22±5.4 | 11±3.2 | 42±3.2 | 47±3.8 |

* clay ≤ 0.002 < silt ≤ 0.05 < sand ≤ 2 mm

Table 4: Arithmetic mean of particle size distribution and bulk density of soils from Pelican Hills and Lac La Biche locations used for measurement of $K_{sat}$ by permeameter

| Horizon | *N* | Particle Size Distribution* Mean % ± 1 *SE* | | | Textural Class | Bulk Density ($Mg\ m^{-3}$) |
|---|---|---|---|---|---|---|
| | | Sand | Silt | Clay | | |
| Ae | 6 | 36±7 | 50±6 | 14±1 | Silt loam | 1.63 |
| Bt | 6 | 32±3 | 26±1 | 42±3 | Clay | 1.55 |

* clay ≤ 0.002 < silt ≤ 0.05 < sand ≤ 2 mm

Table 5: Bulk density of Pelican Hills profiles

| Increment | Forested | | Logged | | 2-tail Sig. |
|---|---|---|---|---|---|
| | Bulk Density ($Mg\ m^{-3}$) Mean % ± 1 *SE* | *N* | Bulk Density ($Mg\ m^{-3}$) Mean % ± 1 *SE* | *N* | |
| LFH (mm) | 0.145±0.004 | 20 | 0.160±0.006 | 16 | 0.057 |
| 0-100 | 1.05±0.03 | 24 | 1.22±0.03 | 12 | 0.001* |
| 100-200 | 1.33±0.02 | 24 | 1.33±0.03 | 14 | 0.820 |
| 200-400 | 1.36±0.04 | 16 | 1.47±0.03 | 8 | 0.093 |
| 400-600 | 1.45±0.02 | 16 | 1.51±0.04 | 8 | 0.210 |

* significant at $\alpha = 0.05$

Table 6: Kostiakov infiltration equation parameters, $R^2$ and sample size at the Pelican Hills site

| Parameter | Ae Horizon | | Bt Horizon | |
|---|---|---|---|---|
| | Logged | Forested | Logged | Forested |
| slope (m) | 0.43⊥ | | 0.54‡ | |
| intercept (a) | -0.31⊥ | -0.013‡ | -0.72⊥ | -0.36‡ |
| Mean $R^2$ | 0.95 | 0.97 | 0.97 | 0.98 |
| Number of Measurement Locations | 23 | 28 | 23 | 27 |

⊥,‡ symbols indicate differences in slope (m) between horizons and differences in intercept within horizons, respectively.

Table 7: Saturated hydraulic conductivity of Gray Luvisolic soil horizons or profiles of the Athabasca and Grandin series as measured by core samples and falling head permeameter or as inferred from measurements of steady-state infiltration flux in double ring infiltrometers

| Method of Measurement | Soil Horizon or profile | Orientation | Geometric mean ($m\ s^{-1}$) | Standard deviation (log units) | *N* |
|---|---|---|---|---|---|
| core + permeameter | Ae | vertical | $5.7 \times 10^{-7}$ | 0.55 | 12 |
| core + permeameter | Ae | horizontal | $1.0 \times 10^{-6}$ | 0.96 | 15 |
| core + permeameter | Bt | vertical | $3.6 \times 10^{-6}$ | 0.87 | 13 |
| double ring | forested | vertical | $5.3 \times 10^{-6}$a | 0.48 | 27 |
| double ring | logged | vertical | $2.2 \times 10^{-6}$b | 0.68 | 23 |

a, b: indicate significant differences among samples obtained by double ring infiltrometer

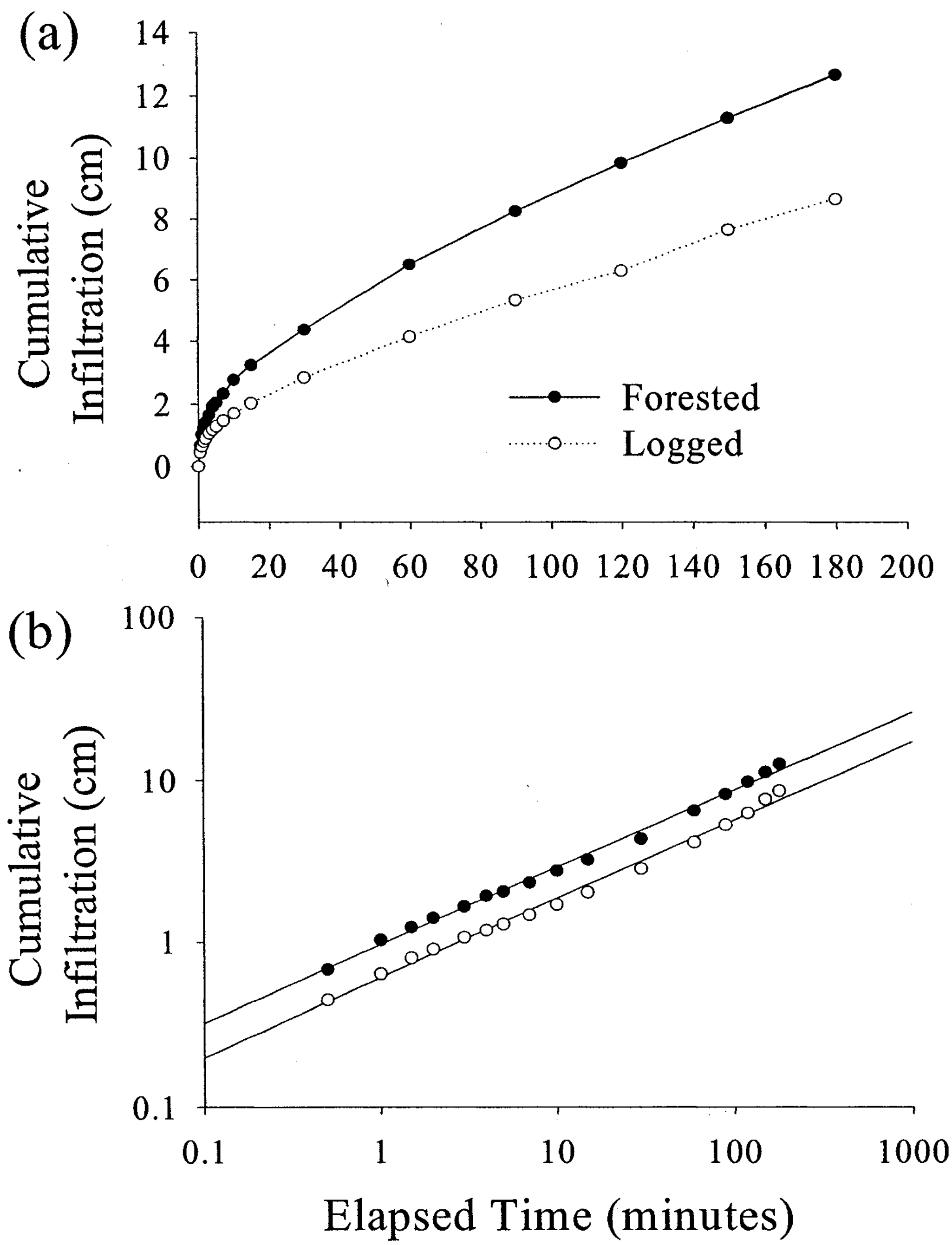


Figure 12: Mean cumulative infiltration to t = 180 through Ae horizon surfaces at Pelican Hills location: (a) linear scale (b) log scale with best-fit line

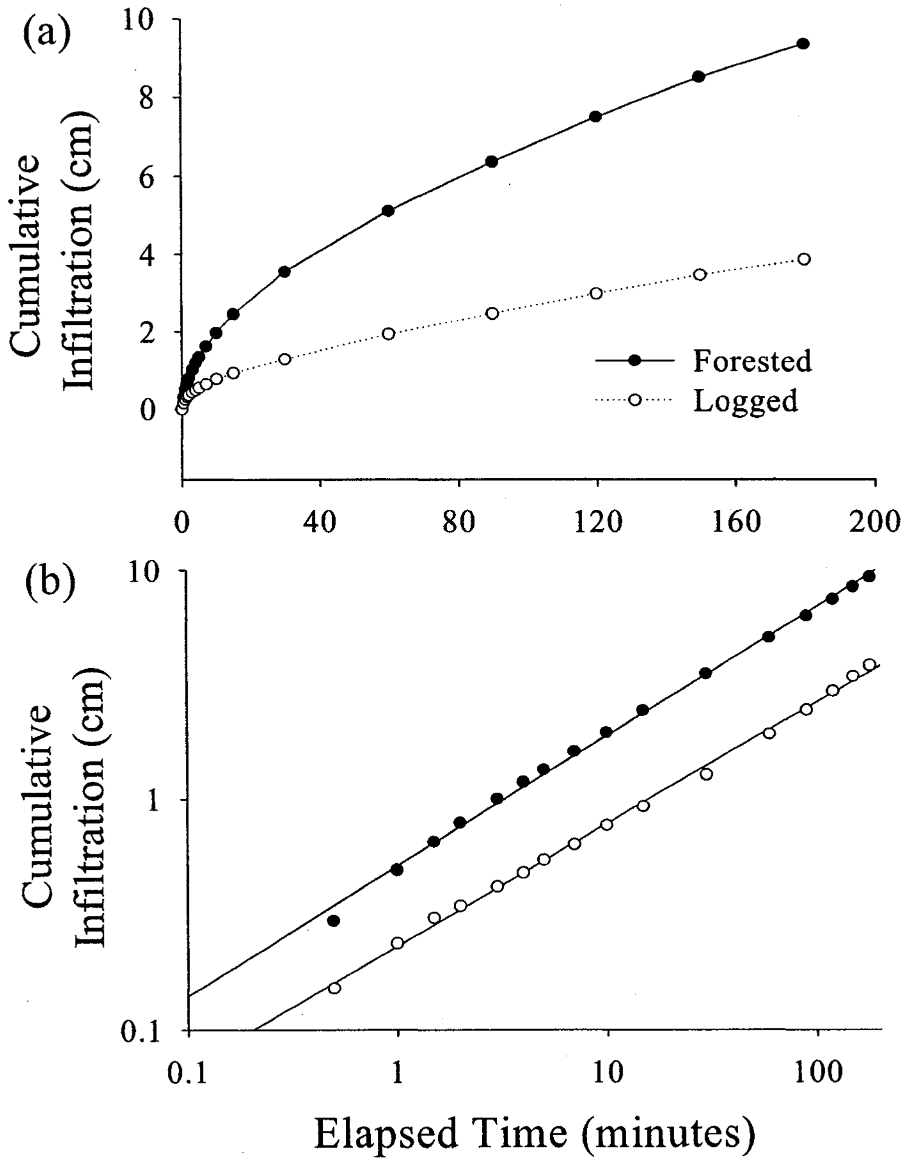


Figure 13: Mean cumulative infiltration to t = 180 through Bt horizon surfaces at Pelican Hills location: (a) linear scale (b) log scale with best-fit line.

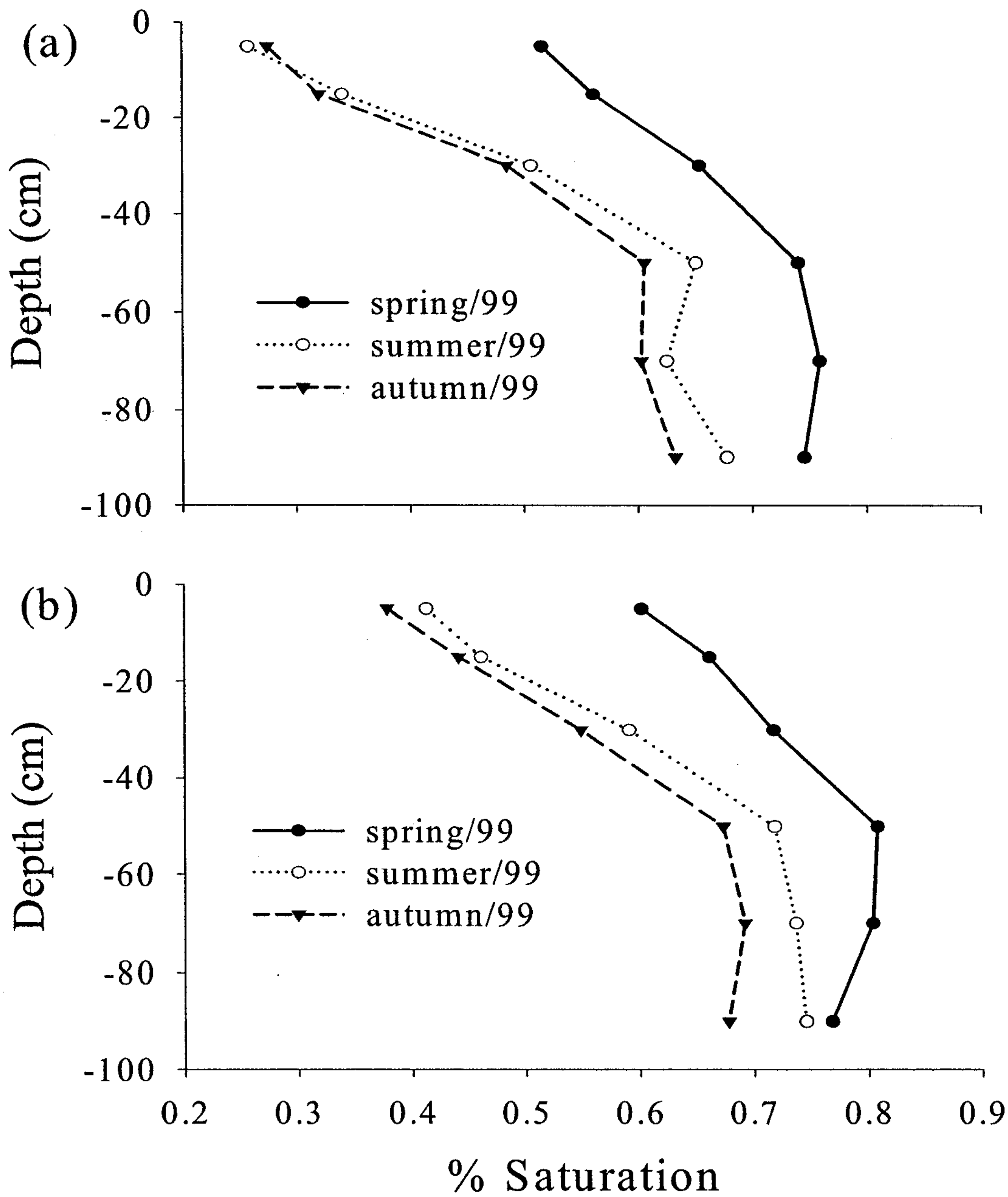


Figure 14: Water content at plots adjacent to sites used for double ring infiltrometer tests at Pelican Hills location: (a) forested (b) logged.

# Chapter 4

## Patterns of water movement on a logged Gray Luvisolic hillslope during the snowmelt period[2]

### *Introduction*

The mechanism of water movement on the hillslope continues to defy simple explanation. Yet understanding how runoff is generated is necessary to build reliable deterministic hydrologic models. It is even more important to correctly identify hydrologic pathways if these models are to accurately simulate solute transport. Views of hillslope flow generation continue to evolve as new knowledge is gained. The mechanism of infiltration excess overland flow (Horton 1933) was supplanted for many forest systems by the concept of the variable source area (Hewlett and Hibbert 1967), as it became clear that streamflow was generated from a limited part of the watershed. Subsurface flow was also recognized, particularly in forest systems (Whipkey 1965; Weyman 1973), for delivering hillslope flow to the stream. Early modeling, though, showed that the hillslope was not always important for stormflow generation due to the low permeability of many soils (Freeze 1972). Processes of saturated overland flow (Dunne and Black 1970), with flow often derived from topographic lows (Anderson and Burt 1978), fit low relief landscapes in particular

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(Anderson and Burt 1990). Evidence supporting rapid hillslope contribution to stormflow though was explained by preferential flow (Beven and Germann 1982). The assumption that water leaving the hillslope in macropores was "new" water has gradually been replaced by a recognition that even in settings where pipeflow is the norm, stormflow has an old water signature (McDonnell 1990). Hillslope flow is currently viewed as involving multiple mechanisms that vary in space and time that are triggered by thresholds (Sidle *et al.* 2000; McDonnell 2003) depending on local soil and weather conditions. On the Boreal Shield Ecozone of eastern North America, for instance, Horton overland flow occurred on rock outcrops whereas subsurface flow occurred in areas of shallow soil (Allan and Roulet 1994). Hillslope flow generation in forest settings often includes interflow in locales of shallow soil cover (Renzetti *et al.* 1992) or deep (Gaskin *et al.* 1989; Mulholland *et al.* 1990; Wilson *et al.* 1990; Wilcox *et al.* 1997). The key requirement is a horizon of reduced hydraulic conductivity (Whipkey and Kirkby 1978; Ahuja and Ross 1983).

The mechanisms of flow generation from hillslopes in the Boreal Plain Ecozone of western Canada, particularly during the snowmelt period when soils are frozen, have received relatively little study in comparison to other regions. Boreal Plain hillslopes are dominated by Gray Luvisolic soils (Figure 1) (Ecological Monitoring and Assessment Network 1996) that occur on common clay-dominated till deposits (Pawluk and Bayrock 1969). Gray Luvisols have fine textured subsoil horizons (Howitt and Pawluk 1985) that restrict percolation (Coen and Wang 1989). Snowmelt is a particularly important period for the hydrologic cycle in the Boreal region of Canada, with 25 to 35% of annual precipitation typically occurring in winter months (Environment Canada 2002a). The contribution to surface waters of the part of the hillslope where Gray Luvisolic soils develop is particularly uncertain relative to contributions from soil types that form at lower slopes (e.g., Organic soils).

The fate of snowmelt depends on vegetation type, soil conditions, and climate, and includes movement into the atmosphere, into soil and ground water storage, and into surface waters such as streams, lakes, and wetlands. Flow from uplands during the thaw period supports the existence of wetlands in both arctic (Young and Woo 2000) and prairie watersheds (Hayashi *et al.* 1998). On the western Canadian prairies, seasonally frozen soils often limit infiltration (Granger *et al.* 1984) and produce snowmelt runoff as infiltration excess overland flow. The snowmelt period is the most important period for annual runoff in prairie (Hayashi *et al.* 1998), montaine (Wilcox *et al.* 1997), and arctic regions (Young and Woo 2000). Kane and Stein (1983) showed that soil frost reduced infiltration capacity of silty textured Alaskan forest soils. In both forest and prairie soils, initial pre-freezing soil moisture content was closely related with the magnitude of infiltration capacity reduction from freezing (Kane and Stein 1983; Granger *et al.* 1984). Frozen soil conditions may increase the contribution of Boreal Plain hillslopes to watershed stream discharge along shallow flowpaths including overland flow and shallow interflow.

Despite demonstrated reductions in infiltration capacity in soil profiles, there is considerable field evidence that soil frost has minimal impact on the hydrologic pathways in forest soils of medium to coarse texture or high organic matter content. In these settings, soil frost does not reduce infiltration capacity enough to generate Horton overland flow in the typically low snowmelt rates typical of northern forests (Price and Hendrie 1983). For a common forest soil catena with coarse textured till soils on upper slopes and organic soils in lower slopes in northern Sweden, snowmelt infiltration raised watertables and thus increased subsurface flow towards the stream (Nyberg *et al.* 2001). Frost had minimal effect on producing shallower subsurface or overland flow, either because soil frost was minimal during years of normal or excessive snow accumulation, or because of reduced snowpack in years when deep frost could develop. Similar soil and climatic conditions to those in the Swedish study were found at the Lac Laflamme watershed in Quebec, Canada. Here, coarse textured soils developed on shallow till deposits, although frozen at the time of snowmelt, had no overland flow and minimal subsurface flow in the vadose zone. Most flow occurred below the permanent watertable, often as pipe flow (Roberge and Plamondon 1987). Price and Hendrie (1983), who worked with coarse textured, homogeneous soil profiles in Ontario, found that vertical rather than lateral flow dominated, with most snowmelt recharging soil moisture or groundwater. Only when a frozen layer developed at the base of the snowpack did Horton overland flow occur. They believed most runoff was produced by saturated overland flow from areas of shallow watertables. Despite frozen soils, groundwater flow was found to be the dominant pathway during the snowmelt period at sites in Manitoba with both fine and coarse textured soils overlying crystalline, Precambrain bedrock (Thorne *et al.* 1998). Infiltration recharged groundwater on upland sites and increased soil water pressure in nearby discharge areas. The findings from these diverse site types suggest that the hydrologic flowpaths and source areas feeding surface waters remain the same regardless of whether soils are frozen or not.

Evidence, however, from some field and modeling studies suggests that frost does influence infiltration and the nature of hydrologic flowpaths. Horton overland flow did occur in forest sites in relatively fine textured soil in New Mexico (Wilcox *et al.* 1997). In this setting, snowmelt runoff occurred in 3 of 4 years, both as surface and subsurface flow, and increased with antecedent moisture content as the model of Granger *et al.* (1984) predicted for fine textured prairie soils. In similar montane settings in Wyoming, Horton overland flow during the snowmelt period only developed when clay content of the soil exceeded 28%. At sites with coarser textured or hydrophobic conditions, snowmelt produced subsurface flow via perched water tables (Nyberg and Fahey 1988). Stahli et al. (2001), who used a model calibrated for Swedish forest soils, showed that flow simulations were more accurate when infiltration capacities were reduced for soil frost. The authors showed that an albeit rare delay in winter snowpack development would allow deep frost penetration and subsequently impact the hydrograph. Similarly, Takata (2002) was better able to simulate changes in soil water storage during the snowmelt if permeability was adjusted for ice content of the soil.

Some of the contrary findings reported in the literature may reflect differences in soil texture, melt intensity, or of the depth and kind of soil frost that develop. Concrete frost is far more effective at reducing infiltration capacity than porous or honeycomb types that tend to occur in forest soils (Price and Hendrie 1983; Jones and Pomeroy 2001). Price and

Hendrie (1983) found only porous frost rather than concrete frost in organic soil horizons at their boreal forest site. Catchment runoff at the Lac Laflamme watershed was related to the spatial extent of concrete frost development (Jones and Pomeroy 2001). While it has been assumed that infiltration is limited to thin unfrozen water films (Kane and Stein 1983), infiltration into air-filled pores has been recognized as another important pathway that allows infiltration capacities of frozen soils to remain greater than melt rates (Stahli *et al.* 1996). Finally, models that are limited to expressing the movement of water in just one and two dimensions may be unable to correctly describe the effect of frost on the three-dimensional movement of water.

Soil frost in Gray Luvisols may result in increased water flow generation relative to the summer months and may involve shallower flowpaths, including interflow through the LFH and Ae horizons or overland flow. There is evidence of frequent annual ice lens formation in the Ae horizons of these soils, resulting in the widespread characteristic of horizontal or platy structural units (St. Arnaud and Whiteside 1964). The platy structure of Luvisolic Ae horizons is most strongly expressed in colder parts of this soils geographic range (Rust 1983). Horizontal ice lenses and evidence of higher lateral than vertical permeability in Ae horizons (Whitson *et al.* 2003a) suggest that infiltration and subsequent downward percolation may be restricted by soil frost, allowing development of perched water tables, and subsequent shallow interflow or saturated overland flow. The additional influence of forest harvesting may further increase flow from hillslopes in this region (Kachanoski and De Jong 1982). Timber harvest under frozen conditions was found to increase bulk density in upper mineral soils and reduce infiltration capacities to approximately 40% of undisturbed values (Whitson *et al.* 2003a).

### *Objectives and Hypotheses*

The objective of this study was to determine the importance of lateral water movement during snowmelt on a Gray Luvisolic soil dominated forest hillslope. I hypothesized that:

- infiltration would be limited by frozen soil conditions and recent forest disturbance, with the result that both overland flow and interflow would occur on the hillslope.

To test these hypotheses, I documented changes in liquid soil water content, soil temperature and measured interflow and overland flow rates along surface and subsurface flowpaths at a recently harvested forest site in the northern Alberta portion of the Boreal Plain Ecozone.

### *Methods*

The study site was located near Smith, Alberta (55°24'35" N, 113°42'37" W), approximately 250 km north of Edmonton, Alberta, Canada (Figure 15). This site was adjacent to one of the 12 study lakes used in the Terrestrial, Riparian, Organisms, Lakes and Streams (TROLS) project that examined relationships between riparian buffers and lake water quality after forest harvest (Prepas *et al.* 2001). The site was chosen because of suitable topography and soils for a study of the movement of dissolved phosphorus after timber harvest. The instrumentation installed for these studies and the support from the TROLS infrastructure allowed us to investigate the processes of water movement during snowmelt. Upland areas prior to harvest were vegetated by aspen (*Populus tremuloides*) and white spruce (*Picea glauca*), typical of the Boreal Mixedwood Ecoregion (Strong and Leggat 1992). Topography was hummocky, and Gray Luvisolic soils dominated the morainal landscape, where drift thickness was forecast to be between 15 and 45 m in thickness (Pawlowicz and Fenton 1995). Clearcut logging took place in November, 1996, after which the harvested areas received no additional silvicultural treatment. The study site was located within the Driftwood River catchment, a 2100-km$^2$ basin that drains the southern slopes of the Pelican Mountains into the Athabasca River near Smith, Alberta. A hillslope with 13% gradient, easterly aspect and convex shape was selected for instrumentation. The hillslope site was approximately 0.5 ha in size within a harvested area of approximately 40 ha and lay on the west side of an adjacent lake. A 20-m buffer of mature forest remained between the lake and the harvested area. Soil profiles at upper, mid, and lower slope positions were similar with respect to thickness, texture and structure and belonged to the Grandin series, an Orthic Gray Luvisol developed on fine-textured till. A typical profile consisted of a 8-cm LFH horizon composed of decayed plant material, a 20-cm thick silt loam textured Ae horizon, underlain in turn by clay-textured Bt, BC and Ck horizons. Water tables were approximately 5 m below the surface at the lower slope position. The great depth of glacial deposits coupled with the great depth to the water table suggests that downward percolation is a distinct possibility on this hillslope. Volumetric water content of Ae and Bt horizons at saturation was estimated from measurements of bulk density. Since harvest the site had naturally revegetated mainly to a very dense stand of aspen that had already attained a height of 2 m. The hillslope site was instrumented in 1998 and measurements were obtained in 1999.

Type T (copper-constantan) thermocouple wire of equal lead length was installed at depths of +5, -10, -25, -45, -65, -85 and -105 cm relative to the mineral soil surface at upper, upper-mid and lower-mid slope positions at each of the three transects (Figure 16). Thermocouples were installed with a hand auger to minimize profile disturbance and auger holes were then backfilled with sand. Temperature was measured manually several times weekly from February to May at various times of the day with a digital thermometer. The type T thermocouple had an accuracy of ±1%, while the thermometer had a resolution of 0.1 °C and an accuracy of ±0.1% of the measured value. The presence of frost was not regularly verified. However, excavation of several soil pits revealed that LFH horizons could be removed by shovel while frozen but that underlying Ae and Bt horizons required heating to first thaw the soil, suggesting that concrete frost was present in the mineral soil horizons.

Liquid soil water content was recorded on a continuous basis with Campbell Scientific CS 615 water content reflectometer probes (TDR) installed horizontally at a 15-cm depth (lower Ae horizon) and at a 40-cm depth (Bt horizon) at the upper-mid slope position of each transect (Figure 16). The TDR probes had an accuracy of ±2% and a resolution of approximately $10^{-6}$ $m^3$ $m^{-3}$ (Campbell Scientific Inc. 1996). TDR probes were installed in holes dug to the appropriate depth. The probes were then laid flat and the holes were backfilled with original soil horizons as closely as possible to their background condition. Even though horizons were replaced in their natural order, installation of TDR probes in this manner probably affected hydrologic characteristics, particularly by changing the platy soil structure of the Ae horizon. The LFH horizon was minimally affected by excavation, so the thermal regime at these locations was not likely changed. Measurements were recorded hourly with a Campbell Scientific CR10X datalogger and converted to volumetric water content after the probes had been calibrated for the Ae and Bt soil horizon materials. Calibration was carried out by adding known quantity of soil from respective horizons to PVC tubes ranging in diameter from 7 to 15 cm so that the entire TDR probe was buried. Soils were brought to saturation by allowing water to fill tubes by capillary suction and also by raising the water table around the tube to equal the soil height. TDR return period was recorded continuously beginning with saturation and allowing several months of drainage and evaporation. Weight of soil was recorded near daily for the first month, then weekly thereafter over a calibration period of approximately 115 days. Soil tubes were heated to 35 $^{o}$C to reduce water content as low as possible near the end of the trial. Two replicates of both Ae and Bt horizon material were used.

Air temperature was recorded hourly with a Campbell Scientific combination relative humidity-temperature probe. The daily maximum hourly-average air temperature was also recorded by datalogger. Soil water content was also measured the previous autumn and again in the spring manually by sampling all slope positions of each transect in increments to 100-cm depth. Water content of soil samples was converted to a depth basis ($m^1$ $m^{-1}$) with bulk density measurements and by integrating volumetric water content over the 1.0-m profile depth. Snow depth and water equivalent were measured at each transect during the late winter of 1999. Snow depth was measured with a metre-stick at approximately 20 points over the hillslope and water content was obtained from 10 points using a plastic coring tube with an inner diameter of 65 mm to obtain snow samples.

Flow was intercepted at three depths in the profile at the lower-midslope position of each transect with pairs of rectangular stainless steel troughs, each 1.22 m in width for a total pair-width of 2.44 m (Figure 16 and 17). Zero-tension collectors in the Ae horizon intercepted the upper and lower 6 cm of the horizon, while the LFH flow collectors enclosed the entire LFH horizon thickness and protruded above to also capture overland flow. Collectors were inserted horizontally upslope from trenches dug to expose soil horizon boundaries (Figure 17). After tubes were attached to the troughs, trenches were backfilled to minimize disturbance of water movement patterns as per Atkinson (1978). Flow was measured continuously with automatic tipping bucket flowmeters beginning at the onset of snowmelt in March (YD 84) and results were recorded hourly. Catchment area was approximated by the 2.44-m collector width and length of slope above the collector. Catchment areas were 112, 110 and 88 m$^2$ for Transects 1, 2 and 3 respectively. Flow intercepted by the uppermost flow collector consisted in principle of both vertical and horizontal components of LFH horizon subsurface flow and overland flow. Since I could not distinguish among the sources of water intercepted by the LFH flow collector, I will henceforth refer to this component as LFH horizon flow. The flow collectors within the Ae horizon on the other hand, intercepted only lateral subsurface flow, or interflow. Thus, LFH flow consists of several possible flow types, whereas flow intercepted in the Ae horizon was strictly interflow.

A soil temperature that stayed near 0 °C over a period of several days was thought to indicate that ice was present in the profile (Thorne *et al.* 1998), as the ice component was expected to dominate the specific bulk heat capacity of the soil (Miller 1980). Once temperature departed on an upward trend from 0 °C, thaw was assumed to have happened. Date of soil thaw was determined by visual inspection of the plot of temperature vs. time.

***Results***

The results of TDR calibration are shown in Figure 18a-b. The relationship between volumetric soil moisture content and signal return period (milliseconds) of the TDR probe was fitted by a two-parameter polynomial equation. Some scatter is evident for the wetter range of values in the Ae horizon, possibly due to soil settling, although the overall regression coefficient ($r^2$) was still 0.95. There was less scatter in the data for the Bt horizon ($r^2 = 0.98$). These calibration curves were superior to the generic relationship provided by the manufacturer of the TDR probes in matching measurements made by gravimetric sampling of soil moisture and estimates of saturation water content obtained from bulk density data. I could not calibrate the TDR for frozen conditions, so the minimum liquid moisture content in my calibration was still probably above that present under frozen conditions. I believe the calibration is valid for frozen soil conditions because the dielectric constant of ice is similar to that of mineral soil. The liquid water component of the bulk soil still has the greatest influence on the effective dielectric constant under frozen conditions.

Antecedent fall soil moisture contents were low, with mean volumetric water content of 18 % in the lower Ae and 23 % in the upper Bt. Winter snowpack remained at less than approximately 15 cm depth of snow until January, when most of the winter snow fell. The mean December temperature at the study site was -14 C. Such conditions of minimal snowpack and cold temperatures would have allowed deep frost penetration. Winter snow fall depth was not recorded, but was less than the long term average depth of 144 cm (Environment Canada 2002a). Snow was evenly distributed over the slope on Year Day (YD) 56 prior to melt, with a mean depth of 48 cm and a standard deviation of just 4 cm. Snow density averaged 0.18 Mg $m^{-3}$ for an average snow water equivalent (SWE) of 87 mm. No further snow fell after YD 56 except on YD 95 to 96 when about 10 cm fell. This latter event included some rain. Water stored in the LFH horizon and upper 100 cm of mineral soil increased from a mean ± *SE* of 296 ± 11 mm on October 22 to 400 ± 7 on May 15 of the year following, for a net increase of 104 mm (Whitson *et al.* 2003b). Between October 1998 and May 1999 soil moisture storage was estimated to have increased from 33 ± 4 to 41 ± 4 mm in the LFH horizon. Thus, less than 10 mm of SWE was needed to fill storage in the LFH horizon.

Average soil temperature at any given depth increment was similar among slope positions and thaw tended to occur at the same time (Figure 19b-h). A soil temperature near 0 °C was assumed to have resulted from frozen conditions (Thorne *et al.* 1998). Based on soil temperature measurements, soil profiles at all slope positions in the late winter to early spring period were probably frozen from the surface to a depth of 65 cm. The resolution of my data are not fine enough to confirm whether the 85 cm increment was frozen, but it is doubtful if soils were frozen at 105 cm depth. The LFH horizon thawed on average on YD 101, and thereafter LFH horizon temperature tended to follow air temperature (Figure 19a), and probably refroze when air temperatures fell below 0 °C. The averge date of thaw for the 10 and 25 cm increments was YD 115, whereas the 45 and 65 increments appeared to thaw near YD 120. Date of thaw ranged from YD 98 to 105 in the LFH horizon and from YD 103 to YD 120 at the 10-cm depth (mid-point of the Ae horizon). Median date of complete profile thaw at the nine measured locations was YD 125 with a range from YD 118 to135 (data not shown).

Visual estimates and photographs of snow cover were used to document the completion of melt on the hillslope over the remaining days. On YD 84, the entire hill was snow-covered. On YD 93, snow still covered approximately 40% of the slope; by YD 94, snow cover had decreased to about 25%. Snow cover was estimated at less than 10% on YD 98, and occupied small patches of less than 1.0 $m^2$. By YD 104 all the snow had melted. Snowmelt was therefore completed over the 21-d period of YD 84 to 104.

Soils were near 0 °C and frozen during the period of snowmelt infiltration. Liquid soil water content measured by TDR probes increased dramatically to a maximum in the Ae and Bt horizons over the period YD 91 to YD 103 (Figure 20a-b). Moisture content began to increase in the Ae on YD 92-93 at Transects 2 and 3, and YD 98 at Transect 1. In the Bt horizon, moisture content began to increase on YD 90 at Transect 3 and on YD 98 at Transects 1 and 2. The finding of earlier increase in the underlying Bt horizon than the overlying Ae horizon at Transect 3 probably reflected the difference in location of the two probes, since the Bt horizon TDR probes were not directly below those of the Ae, but rather were approximately 0.5 m to one side. The period of rapid moisture rise in the Ae horizon was 55 h (Transect 1), 125 h (Transect 2), and 140 h (Transect 3) (Fig 6a). Moisture content rose in the Bt horizon for 101 h (Transect 1), 37 h (Transect 2), and 76 h (Transect 3) (Figure 20b). Liquid soil water content fluctuated in the Ae horizon at Transects 1 and 3 from YD 98 to YD 103. Over this period of cyclic behavior, maximum moisture content at Transect 1was recorded at 4:00, 23:00, and 21:00 h whereas moisture content at the minimum in the cycle was recorded at 14:00, 11:00, and 10:00 h. At Transect 3, maximum moisture content was found at 18:00, 14:00, 19:00, 17:00, and 16:00 h, while minimum values occurred at 7:00, 14:00, 11:00, 10:00, and 10:00 h. The tendency for liquid water content to be highest in the evening and lowest in the morning suggests these fluctuations were caused by freezing and thawing of soil water in response to the changing rate and direction of surface energy flux. Liquid water content in the Ae horizon peaked at different times and maxima at each transect. Transects 1 to 3 reached a maximum on YD 110 at 22:00 h (0.47 $m^1\ m^{-3}$), YD 112 at 12:00 h (0.44$m^1\ m^{-3}$) and YD 100 at 01:00 h (0.40$m^1\ m^{-3}$), respectively. Soil moisture content trends in the 3 Bt horizon measurement points varied differently after reaching their initial maxima. Soil water content of the Bt horizon reached a peak of 0.40 $m^1\ m^{-1}$ at 20:00 h on YD 102 at Transect 1, and then decreased to 0.33 $m^1\ m^{-1}$ at 12:00 h on YD 106, and finally peaked again at 0.39 $m^1\ m^{-1}$ at 23:00 h on YD 116. At

Transect 2, liquid soil water content reached 0.44 $m^1$ $m^{-1}$ at 02:00 h on YD 100, while at Transect 3 moisture content peaked at 0.44 at 22:00 h on YD 93. Moisture content remained nearly constant from YD 100 to YD 113 at transect 2, while at transect 3 moisture content declined gradually from YD 94 to 112. Moisture content of the Bt horizons showed a sudden drop at all 3 transects between YD 112 and 115, probably corresponding to soil thaw. At Transect 1, moisture content declined sharply at 15:00 h on YD 115, while at Transect 2 a sharp decline began at 14:00 h on YD 113. Moisture content declined significantly starting at 12:00 h on YD 112 at Transect 3.

The amount and timing of flow intercepted by collectors varied spatially (Figure 21-23). Soil water had to be at atmospheric pressure to enter the zero-tension flow collectors, so it is likely that the development of local saturated conditions allowed the flow events observed. Amounts intercepted were very small in relation to the amount of SWE and took place intermittently. Flow was first recorded on YD 84 at both Transects 1 and 2. At Transect 1, LFH horizon flow and upper Ae horizon interflow both peaked on YD 98, the latter coinciding with increases in soil moisture in the Ae horizon (Figure 21a-b). Interflow in the lower Ae horizon reached a maximum on YD 116 (Figure 21c), coinciding with soil thaw in the 10, 25, and 45 cm increments. At Transect 2, both LFH horizon flow and upper Ae horizon interflow peaked on YD 84 (Figure 22a-b), before soil moisture had increased at upslope TDR probes. The lower Ae horizon interflow peaked on YD 93 (Figure 22c), coinciding with moisture content increases recorded upslope. LFH horizon flow at Transect 3 peaked on YD 106, while upper Ae horizon interflow peaked earlier on YD 101 (Figure

23a-b), near the time that moisture content reached a maximum in the Ae horizon upslope. Interflow in the lower Ae horizon peaked on YD 93 (Figure 23c), just as soil moisture content was beginning to increase upslope. Interflow in the lower Ae horizon occurred several days later than LFH horizon flow in all 3 transects and than upper Ae horizon interflow at Transects 1 and 2 (Figure 21-23). At Transect 3, however, lower Ae horizon interflow preceded interflow in the upper Ae horizon.

Peak rates of LFH horizon flow and interflow occurred in the warm part of the day during the period of melt up to YD 108. Time of peak LFH horizon water movement ranged from 1300 h (local standard time) at Transect 1 to 1900 h at Transect 3. Time of peak flow in the upper Ae horizon flow collectors ranged from 1600 to 1800 h, while in the lower Ae horizon flow peaked at 1900 h at the two locations where it happened (data not shown). Other than the interflow event for YD 116-120 in the lower Ae horizon at Transect 1, all flow had ceased by YD 108, 4 d after snow disappearance.

The largest interflow event from each transect is plotted in Figure 24a-c. The event at Transect 1 began at 02:00 h on YD 116 and continued until 24:00 h of YD 120 (Fig 24a), with peak flow reaching about 0.5 L $h^{-1}$, for a total flux of 18.8 L. Most of the interflow collected on the hillslope during the snowmelt period was obtained from this event (Table 8), which coincided with soil thaw of the Ae horizon and was preceded by a day when maximum air temperature reached 21.2° C. Snow was absent from the hillslope for almost 2 weeks prior to the onset of this interflow event. High soil moisture content may explain the larger volume of flow at this location. Soil moisture content further upslope from the flow collectors was greater at Transect 1 than at Transects 2 and 3, suggesting that this part of the hill was wetter and therefore more likely to be at saturation than elsewhere. Another event occurred in the lower Ae horizon on YD 93-94 at Transect 2 (Figure 24b) at the same time that soil moisture at the upper midslope position increased from 0.14 to 0.25 $m^1 m^{-1}$. Much of this event occurred from 19:00 to 22:00 h on YD 93. The other notable interflow event occurred in the lower Ae horizon at Transect 3 on YD 92-93. Two hydrographs are evident, both occurring in the late afternoon of YD 92 and 93 respectively (Figure 24c).

Volumes LFH horizon flow and interflow intercepted ranged from 0.5 to 2.3 L in the LFH, 0.25 to 0.6 L in the upper Ae horizon, and 1.05 to 18.8 L in the lower Ae horizon (Table 8). A total of 22.4 L of water was intercepted as interflow by the upper and lower Ae flow collectors at the 3 transects over the entire melt period, with 84% of this from YD 116 to YD 120 in the lower Ae horizon at Transect 1. This pulse of interflow (18.8 L) was approximately 8 times greater than the quantity up to the onset of that event. The very low volumes obtained from the LFH flow collectors while snow was melting are partially due to instrument error, as I believe that LFH collectors were not measuring all possible flow. Based on measured SWE and changes in moisture content of the LFH horizon between fall and spring, 10.6 L of snowmelt should have been intercepted at each collector and approximately 2 L of snowmelt should have been stored in the LFH. The other 8.6 L would have been intercepted, and clearly was not, probably because the radius of the exit tube (< 1 cm) was too narrow to accommodate the flow rate from this horizon. Measurements of LFH horizon flow are therefore more useful for illustrating timing rather than magnitude of flow associated with this horizon.

When expressed on an areal basis, snowmelt interflow on the hillslope was very close to zero. The depth equivalent of interflow intercepted by the lower and upper Ae flow collectors was 0.064 mm and yielded a snowmelt runoff coefficient of essentially zero (0.00077). Some snow could have sublimated or evaporated during the melt period. However, average soil water content of the upper 1.0 m of the profile increased by 104 mm depth equivalent between October 1998 and May 1999, suggesting that most of the water in the snowpack infiltrated(Whitson *et al.* 2003b).

Snowmelt generated more streamflow for the Driftwood River catchment in 1999 than was typical. An average of only 28% of annual streamflow occurred over the April to May period from 1970 to 1990. In 1999, some 64% of annual flow occurred during this period. Annual discharge in 1999 was just 11 mm as compared to the long term average of 119 mm, reflecting low precipitation. Therefore streamflow reflected the importance of snow inputs to catchment soils during this unusually dry period.

***Discussion***

Infiltration capacities at the site, despite recent forest harvest, were sufficient to allow snowmelt infiltration without apparent generation of overland flow. Similar findings were reported by Carey and Woo (1998), who found that all snowmelt infiltrated seasonally frozen soils under an aspen forest in the Yukon territory of northern Canada. Given that soil temperature in the upper profile was near 0 °C during the period of snowmelt infiltration, it appears that some soil porosity was ice-filled. Likewise, the rapid increase in liquid water content in the Bt horizon while soil temperature in the 25 to 65 cm zone was near 0 °C indicated that water percolated through partially frozen subsoil. Steady state infiltration rates through Ae horizons measured in the summer with double ring infiltrometers in nearby harvested locations averaged 7.9 mm $h^{-1}$ (Whitson *et al.* 2003a). If we assume that the time required for the entire 87 mm of SWE to infiltrate was equal to the time of soil moisture increase as shown by the TDR probes, then we can approximate the actual infiltration rate for the site. The average time of moisture content increase in the Ae horizon during the snowmelt was 107 h (range 55-140 h), which means the average infiltration rate was 0.8 mm $h^{-1}$, ten times lower than summer infiltration capacity. Variation in the timing of soil moisture increases and subsequent interflow events were probably due to slight differences in aspect among transects and differences in incident solar radiation due to the shading by the young aspen regeneration and the larger white spruce trees remaining along the east edge of the slope. Soils probably stored snowmelt through some combination of freezing and capillarity. The approximately diurnal cycling of liquid moisture content in the Ae horizon for instance suggests that some of the meltwater refroze after infiltrating. The sites that have reported overland flow during the snowmelt (e.g. Wilcox et al. (1997) or Nyberg and Fahey (1988)) lacked LFH horizons, and had organic matter contents in surface horizons ranging from 6-9% (Knight *et al.*

1985). Frozen infiltration capacities in these montane forest settings may not be comparable to that in upland Boreal soils where LFH horizons are the norm. With snowmelt rates typically 5 mm $h^{-1}$ or less (Price and Hendrie 1983) and high liquid soil water contents of organic soil horizons when frozen (Nyberg *et al.* 2001), sites with LFH horizons less likely to generate overland flow than in settings with surface mineral soil horizons.

The hypothesis of shallower interflow was not supported. LFH horizon flow and interflow through the upper Ae horizon were low during the two-week period when the underlying mineral soil horizons remained near 0 °C. The largest interflow events occurred in the lower Ae horizon, as expected for non-frozen conditions, and not higher up in the profile. Therefore, it appears that interflow in Gray Luvisols occurs in the same region of the profile all year.

Very little interflow occurred, suggesting that soil water recharge, percolation, and evaporation was the fate of most snowmelt. Well drained soils at equivalent slope positions in a morainal, prairie landscape in Saskatchewan allowed percolation to 2.5 m depth (Zebarth and De Jong 1989). The amount of flow measured at this site is low in comparison with those from other studies of snowmelt in western North America. Kachanoski and DeJong (1982) measured 61 mm overland flow from 200 mm snow water equivalent on logged terrain in northern Saskatchewan, Canada. Much of the 3 to 11% of annual precipitation that was converted to interflow and overland flow over a 4-year period in an undisturbed, semi-arid Montane region with fine-textured soils in New Mexico, U.S., was associated with snowmelt (Wilcox *et al.* 1997). Findings from these other locations may not apply fully because of differences in bedrock geology, soil properties, and climate. I have only one year of data, and furthermore, low melt intensity at the site could have reduced flow. Spatially unequal rates of interflow across the hillslope during snowmelt, despite even snow distribution, may be common. Harr, (1977) for instance, found that saturated zones where interflow arose were distributed in a patchwork fashion on a Pacific northwestern US hillslope. Likewise, Woods and Rowe (1996) found a great deal of spatial and temporal variability in the amount of subsurface flow on forested hillslopes in New Zealand. The greater interflow at Transect 1 may reflect the relative position of the wetting front in relation to the time of soil thaw. Soil water content rose later at this transect than at the others, such that wetting front penetration was possibly closer in time to soil thaw than at Transects 2 or 3. The more northerly aspect of Transect 1 could have delayed snowmelt more at the other two transects, although records of snow disappearance are too imprecise to reconstruct spatial melt rates among the transects.

There are several explanations for the lack of flow. The first is that Gray Luvisolic soils produce little flow because of typically high storage capacity and infiltration capacities sufficient to accommodate snowmelt rates. This study was completed two years following harvest, so that aspen vegetation regrowth coupled with reductions in bulk density may have improved soil infiltration capacity and subsurface permeability. The easterly aspect at this site may also have dampened flow response through reductions in peak discharge rates, given that the slope was not under direct sunlight during the warmest part of the day, and snowmelt rates would have been lower than on other aspects. Finally, dry pre-winter moisture conditions reduced the potential for ice blockage in larger soil pores and subsequent reductions in infiltration capacity.

The majority of 1999 streamflow in the Driftwood catchment was associated with snowmelt, probably because summer precipitation was much below normal for that year. The April and May catchment streamflow of just 7 mm was 64% of the annual total. Typically streamflow in this region reaches a maximum in summer due to high June to July precipitation (Environment Canada 2002b). Mean May to July precipitation at 7 northern Alberta, Environment Canada, weather monitoring stations for the period 1961 to 90 was approximately double the amount of rain for the August to September period (Environment Canada 2002a). Snowmelt may recharge soil water to allow subsequent stormflow generation during summer precipitation. The absence of a significant winter snowpack will then mean drier soil post-melt and reduced streamflow response to summer precipitation. Interflow on Luvisolic-dominated hillslopes seems more likely to occur intermittently at isolated locations where soil and moisture conditions permit rather than as a spatially and temporally uniform process over the entire hillslope, although the latter is possible under extreme precipitation events.

### *Conclusions*

Gray Luvisolic soils on a morainal hillslope remained near 0 °C during and for two weeks after snowmelt. Despite forest removal two-years previous, the infiltration capacity of upper horizons remained high enough to prevent overland flow and allow nearly all 87 mm of SWE to infiltrate while soils were frozen. Moisture content in the Ae horizon increased to maximum over a time period that ranged from 55 to 140 hours. Less than 0.1 mm of SWE became interflow, with most occurring through the lower Ae. The lack of flow through the LFH or the upper Ae horizon allows us to conclude that frost had little effect on the hydrologic flowpath. Interflow was spatially and temporally variable on the hillslope, occurring when soil moisture content was increasing rapidly. A single interflow event contributed 84% of the total amount during snowmelt, and happened at the time of soil thaw. Low pre-winter soil moisture content probably contributed both to the low volumes of interflow and apparent high infiltration capacity observed at the site.

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Table 8: Water volume intercepted by flow collectors (L) over the period YD 80 to 130

| Transect | Horizon | | |
|---|---|---|---|
| | LFH | Upper Ae | Lower Ae |
| 1 | 2.3 | 0.25 | 18.8 |
| 2 | 1.3 | 0.6 | 1.05 |
| 3 | 0.5 | 0.45 | 1.25 |
| Total | 4.1 | 1.3 | 21.1 |

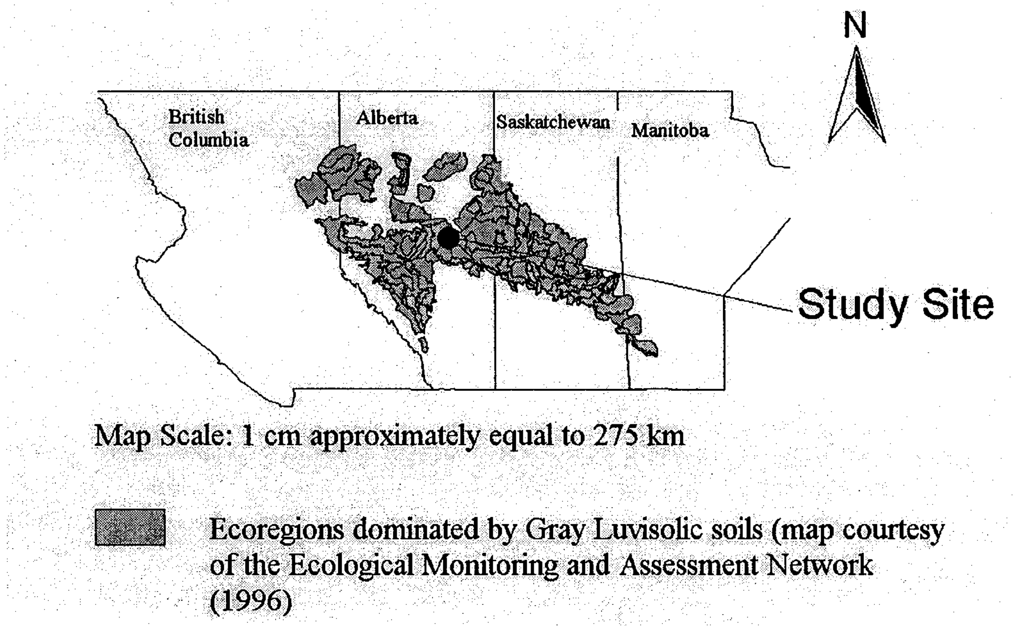
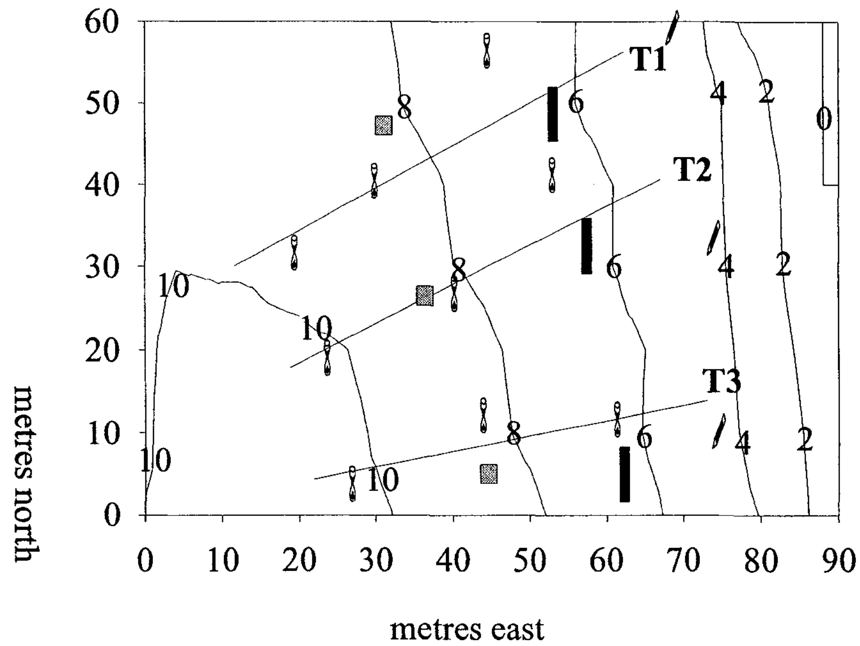



Figure 15: Location of study site including map of ecoregions dominated by Gray Luvisolic soils (map courtesy of the Ecological Monitoring and Assessment Network (1996)




TDR Probes

Subsurface Flow Collector

Groundwater Well

Thermocouple Nest

4 Relative Elevation

T1 Transect

Figure 16: Diagram of study hillslope showing transects, measurement locations and contours with 2 m interval

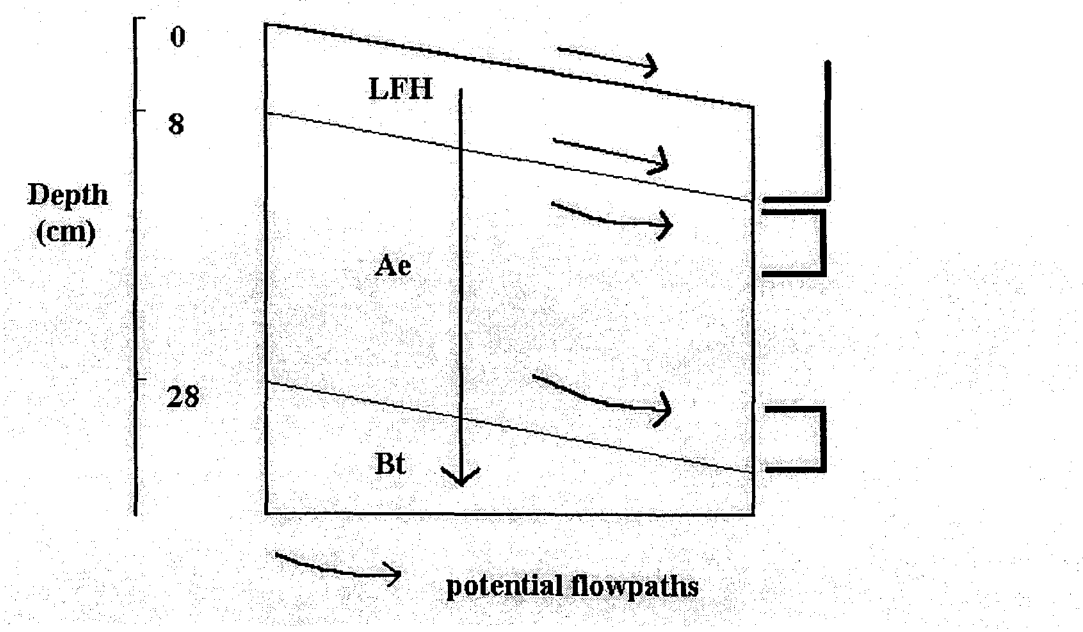



Figure 17: Luvisolic soil profile with horizons, horizon depth in cm, potential flowpaths and flow collectors

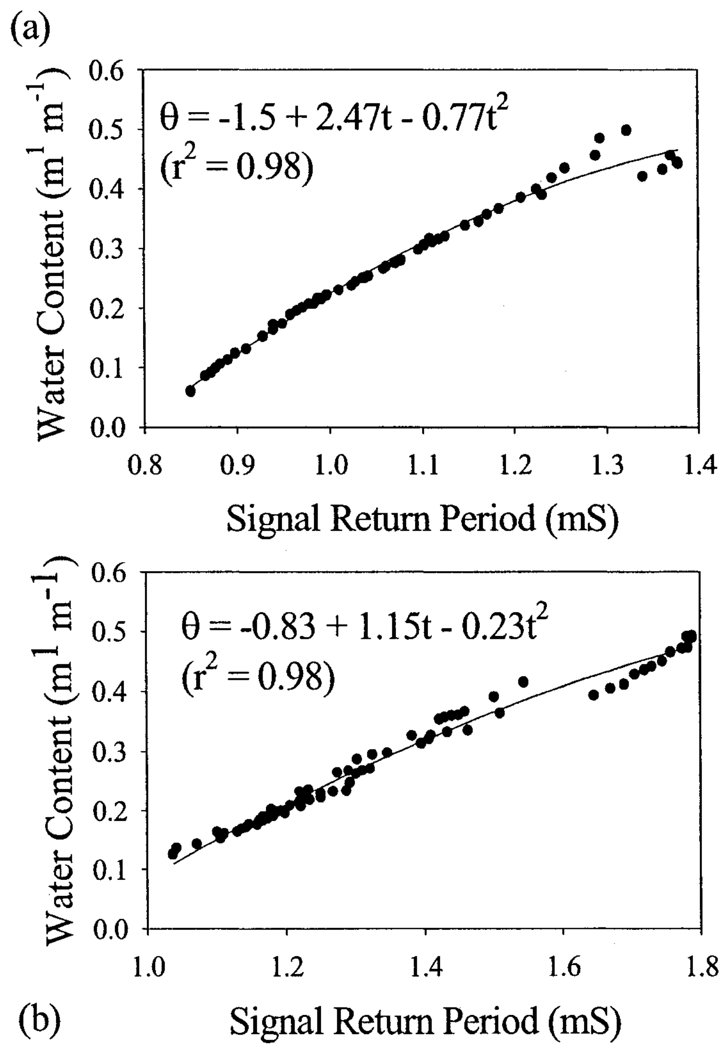
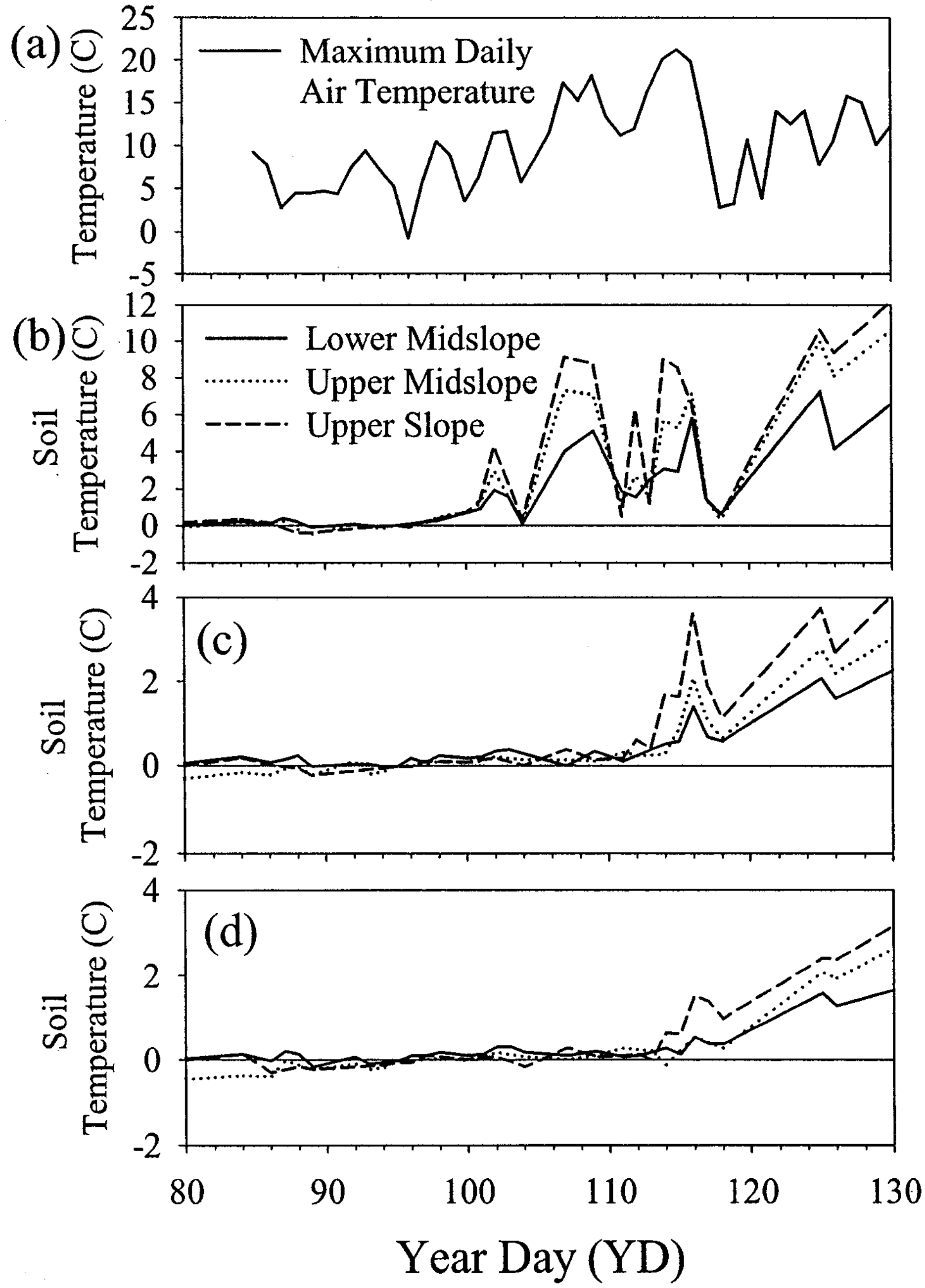



Figure 18: Results of Time Domain Reflectometry (TDR) calibration (a) Ae horizon and (b) Bt horizon



(a)
Temperature (C)
25
20
15
10
5
0
-5
Maximum Daily
Air Temperature
(b)
Soil
Temperature (C)
12
10
8
6
4
2
0
-2
Lower Midslope
Upper Midslope
Upper Slope
(c)
Soil
Temperature (C)
4
2
0
-2
(d)
Soil
Temperature (C)
4
2
0
-2
80
90
100
110
120
130
Year Day (YD)

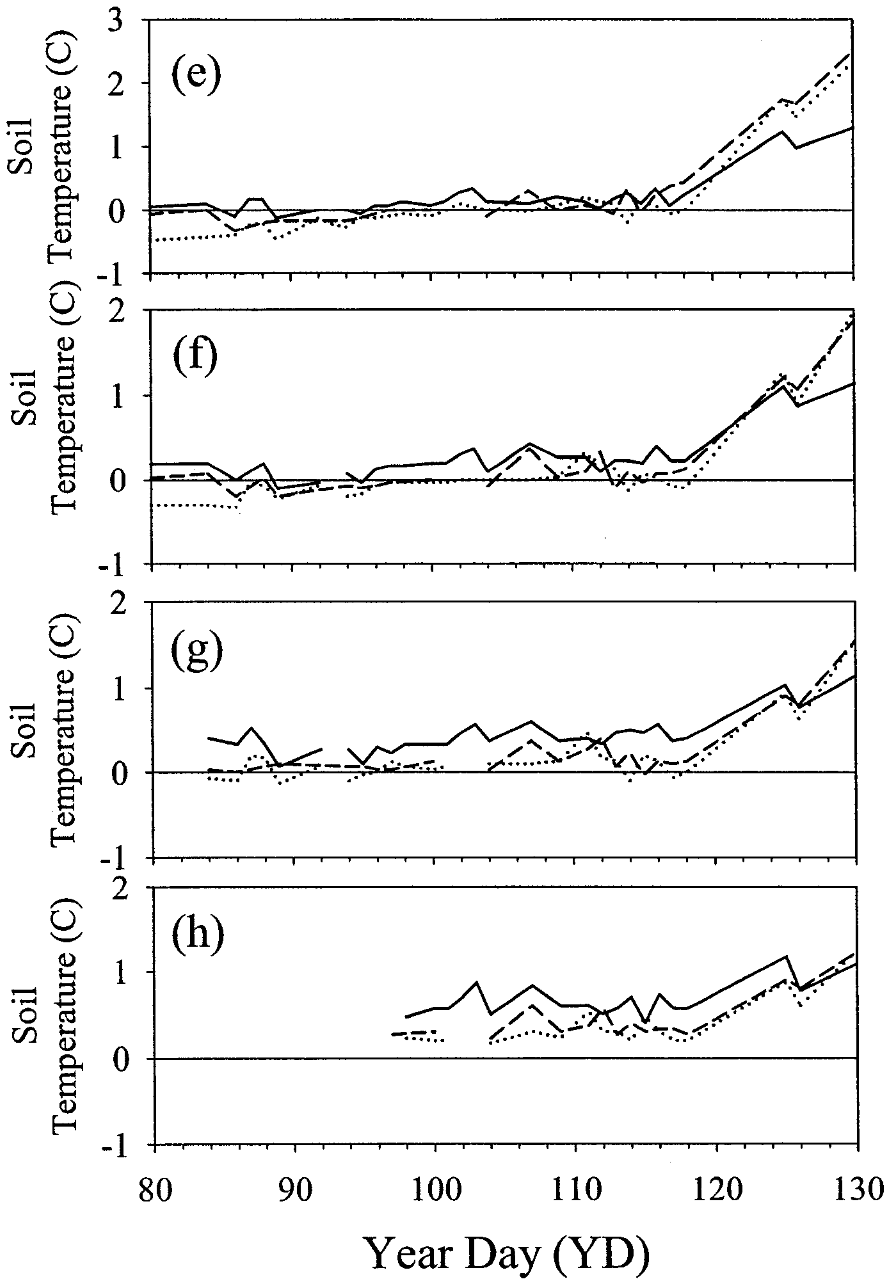



Figure 19: Temperature trends during snowmelt period (a) maximum air temperature (b-h) soil temperature by increment (all depths relative to mineral soil surface) for (b) LFH horizon (c) -10 cm (d) -25 cm (e) -45 cm (f) -65 cm (g) -85 cm (h) -105 cm

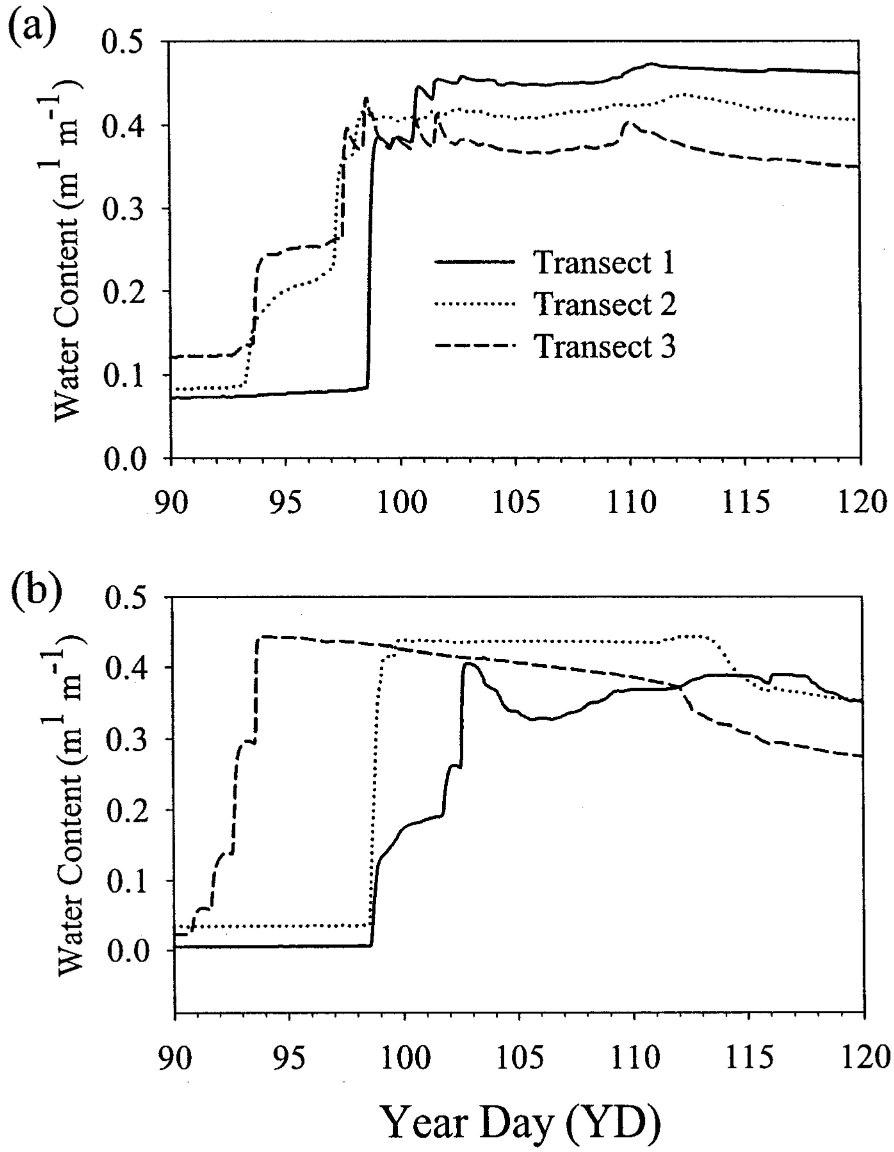



Figure 20: Volumetric water content ($m^1$ $m^{-1}$) by transect for (a) Ae horizon (15 cm depth) and (b) Bt horizon (40 cm depth)

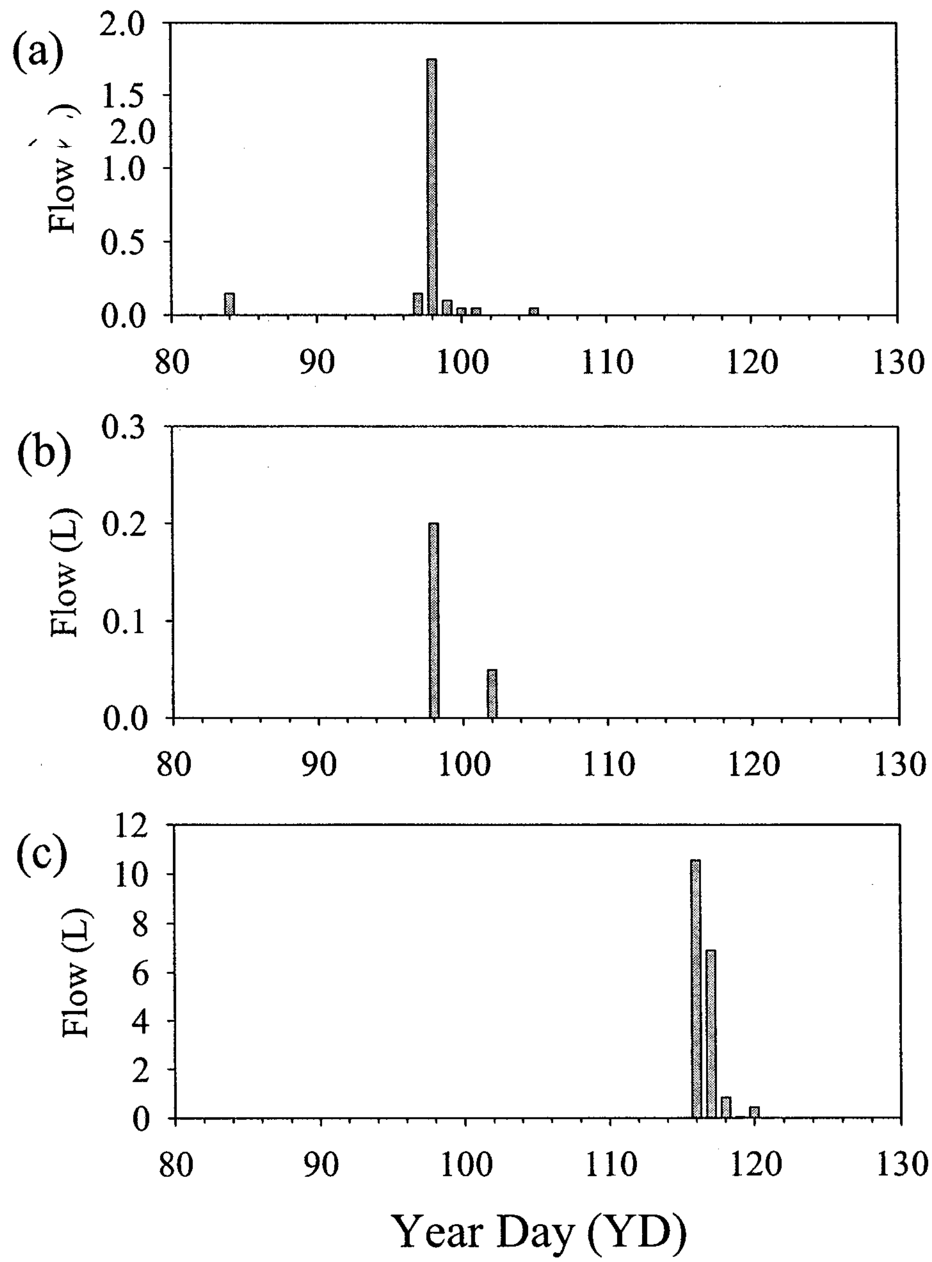



Figure 21: Daily flow obtained by collectors at Transect 1 (a) LFH horizon flow (b) Upper Ae interflow (c) Lower Ae interflow

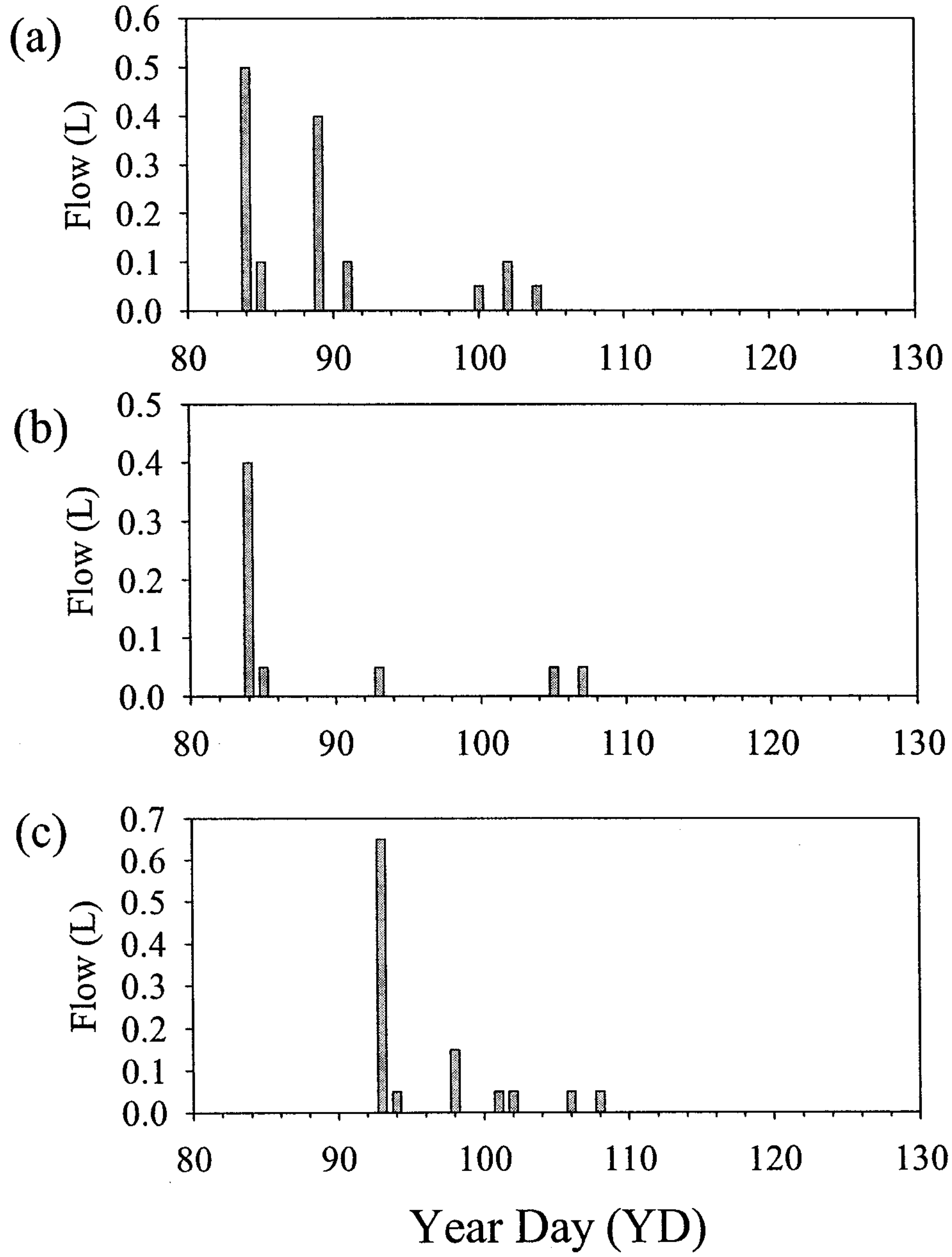



Figure 22: Daily flow obtained by collectors at Transect 2 (a) LFH horizon flow (b) Upper Ae interflow (c) Lower Ae interflow

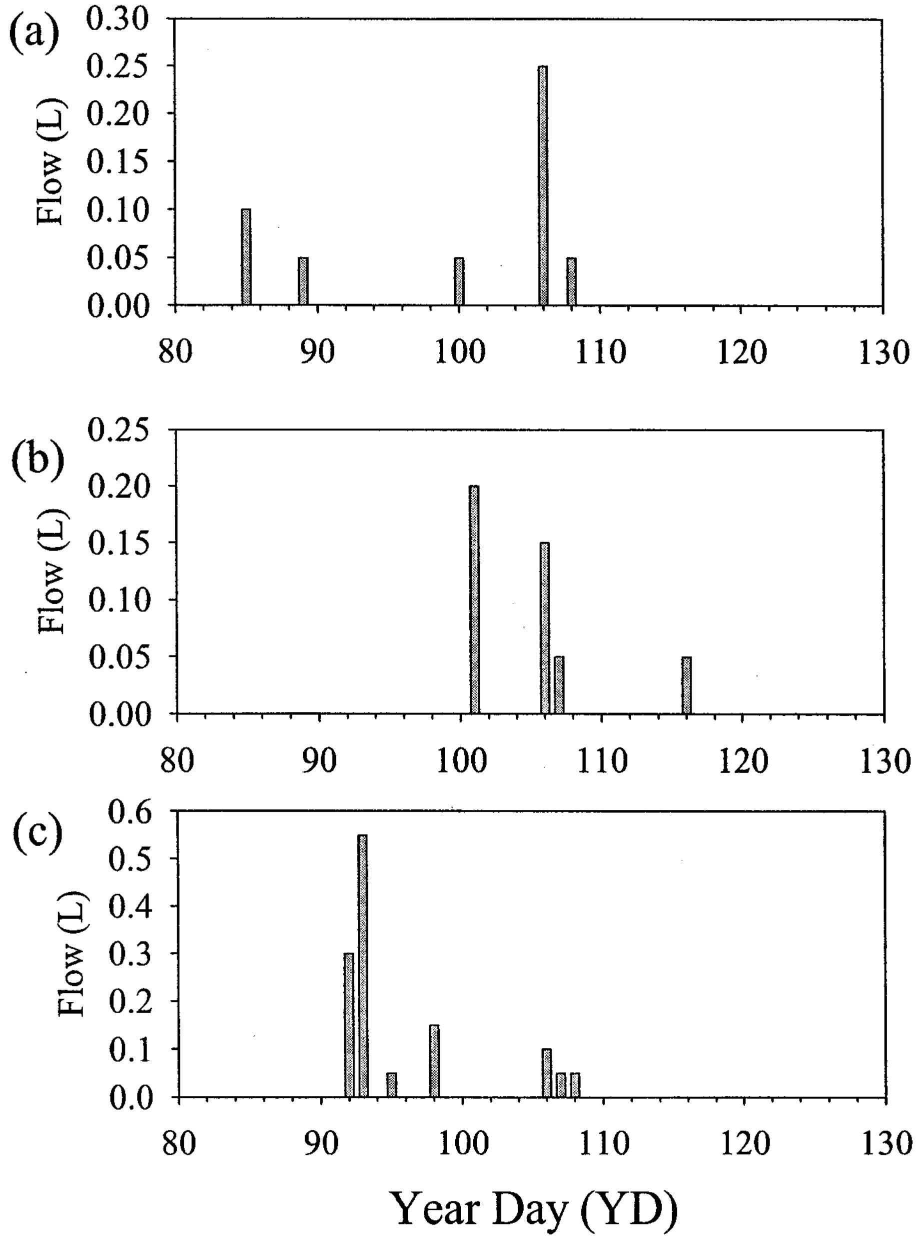



Figure 23: Daily flow obtained by collectors at Transect 3 (a) LFH horizon flow (b) Upper Ae interflow (c) Lower Ae interflow

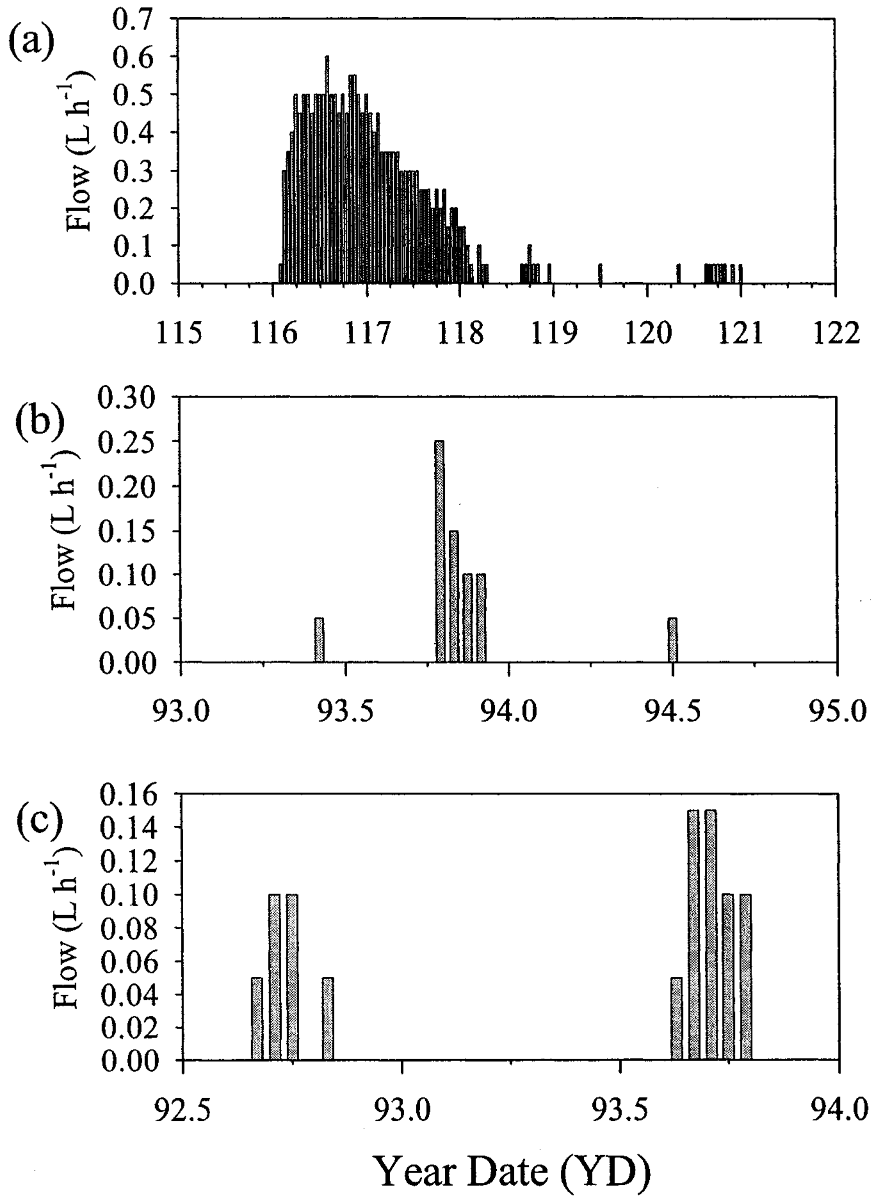



Figure 24: Hourly flow of larger interflow events observed in lower Ae horizon (a) Transect 1, YD 116-120 (b) Transect 2, YD 93-94 (c) Transect 3, YD 92-93

## Chapter 5

## Trends in dissolved phosphorus in Gray Luvisolic soil profiles on a logged hillslope

### *Introduction*

Gray Luvisols have lost more phosphorus (P) since the end of the Wisconsin glacial period than Podzolic soils in eastern North America, possibly through transport with interflow in the A horizon (Xiao *et al.* 1991) or because of enhanced mobility of organic P compounds in the soil solution (Frossard *et al.* 1989; Donald *et al.* 1993). P enrichment of fulvic and humic acid fractions in subsoils confirm the importance of downward movement (Schoenau and Bettany 1987) while downslope movement to lower slope positions along catennary sequences has also been inferred (Huang and Schoenau 1996). Gray Luvisols dominate upland areas of the Boreal Plain of western Canada (Ecological Stratification Working Group 1996). These slightly weathered, young soils are modified most visibly by carbonate removal and clay translocation, and are typically associated with moraines and glacial lacustrine landforms (Ecological Stratification Working Group 1996). Gray Luvisols are typically medium to fine textured and possess mixed mineralogy (Pawluk 1961; Kodama 1979; Abder-Ruhman 1980; Spiers 1982; Howitt and Pawluk 1985a) with montmorillonite dominating the clay fraction. The vertical sequence of typical soil horizons vary predictably in terms of pH, CEC, texture, and organic carbon with depth. The solubility of P will also reflect this vertical gradation in soil properties and controls the amount of P in solution available for transport both above and below the water table (See Table 9 for explanation of terms). Higher P leaching rates suggest that the stronger adsorption reactions that have limited P loss in other forest environments (Wood *et al.* 1984; Yanai 1991) may not be present in Gray Luvisols. In general, Bt horizons sorb less P because of lower amorphous aluminum or iron content than is found in Bf, Bhf, and Bh horizons of acidic Spodosols and Podzols (Ryden and Pratt 1980; Zhou *et al.* 1997). Conversely, organic forms of P may be more mobile than inorganic orthophosphate (Frossard *et al.* 1989). It is important to understand the influence of the soil profile on P movement within hillslopes. This in turn helps account for biogeochemical cycling at larger scales (e.g., movement from uplands to peatlands), for the management and conservation of aquatic ecosystems, which are often P limited (Prepas and Trew 1983) and for quantifying the effect of forest disturbance on P export rates (Prepas *et al.* 2001).

Pedogenesis is accompanied by gradual transformation of solid phase P from primary forms to secondary and organic forms (Walker and Syers 1976). In general, calcium phosphates (fluoroapatite) dominate at higher pH, with iron (strengite) and aluminum phosphates (variscite) becoming more prevalent in acidic profiles (Barber 1995). On the Boreal Plain ecozone of western Canada, this transition from primary to secondary P has begun. Much of the acid extractable P (apatite) in the Ae and Bt horizons of Orthic Gray Luvisols at sites in northern Saskatchewan had been converted to NaOH extractable P (secondary, organic) (Schoenau *et al.* 1989). Similar conversions have been reported for Brunisolic soils in northern Alberta (Fyles and McGill 1988) and reflect the instability of apatite in the typically mildly acidic conditions of the developing solumn. Near complete conversion of apatite in upper horizons of Gray Luvisols has also been reported by Pawluk (1961). The solubility of strengite, variscite and calcium phosphates including apatite reach a maximum at approximately pH 5 (Lindsay 1979). Weathering reactions in upper soil horizons thereby produce conditions most favorable for solid phase P dissolution early in the lifetime of a soil.

Soil organisms also influence P cycling. Many of the organic P compounds associated with organisms (e.g., sugar phosphates) are highly mobile (Rolston *et al.* 1975). Phosphatase enzyme production enhances the conversion of primary to secondary, organic, and soluble P forms (McGill and Cole 1981). Organic acids produced during decomposition can increase P solubility through complexation reactions with metals (Schnitzer 1969; Arp and Meyer 1985; Fox 1993). Changes to the soil moisture or temperature regime following forest disturbance such as fire or logging can in turn affect organisms and their role in P cycling. Overall, soil organisms including fungi accelerate the conversion of organic phosphorus to phosphate.

Concurrent with biological influences are the chemical reactions that limit the concentration of P in solution: adsorption and precipitation. While it is difficult to assign either mechanism to the loss of P from solution (Sposito 1984), the two reaction types can be viewed as operating at different time scales and concentrations. Sorption reactions are important over short time steps and at low concentrations, while precipitation reactions will tend to prevail at higher P concentrations and over longer time periods (Ryden and Pratt 1980). In most soils, adsorption capacities tend to correlate with Fe, Al, organic carbon, or clay (Parfitt 1978). Of these, clay is likely to be most important in forested Gray Luvisols. Both aluminum and organic matter were correlated with adsorption capacity in Gray Luvisols in northwestern Alberta used for agriculture (Soon 1991). Highest iron oxide contents of Gray Luvisols have been found to occur within B horizons (Pawluk 1961). The mechanism of P adsorption on soil particle surfaces usually involves replacement of one to two hydroxyl groups with an orthophosphate oxygen. Hence, the density of these exposed hydroxyl groups determines the capacity and binding strength for P (Ryden and Pratt 1980), and explains the very strong P retention shown by amorphous Fe and Al oxyhydroxides. Movement of clay to subsoils then should reduce P sorption in the eluviated horizon, however, changes to the organic carbon or amorphous iron and aluminum content may counteract the effect of clay movement. Precipitation reactions require the ion activity product to exceed the solubility product of a potential solid phase metal-phosphate. Typically low ionic strength in the soil solution of Gray Luvisols means that precipitation reactions are unlikely.

Soil solution P concentration data obtained in regions with similar soil properties suggest that potential P mobility is spatially variable. Timmons *et al.* (1977) reported flow weighted mean orthophosphate concentrations typically 10 $\mu g\ L^{-1}$ and less in Ae horizons of Boralfs under birch vegetation in northern Minnesota. Soon (1991) reported mean equilibrium P concentrations of 30 $\mu g\ L^{-1}$ in topsoils and of less than 20 $\mu g\ L^{-1}$ in subsoils of agricultural Luvisols in NW Alberta. Shaw *et al.* (1990) found mean concentrations of 21 $\mu g\ L^{-1}$ for total dissolved P in groundwater in the aspen-dominated catchment of Narrow Lake, Alberta. Concentrations of P in leaf litter leachate ranged from 100 to 250 $\mu g\ L^{-1}$ during the snowmelt period in Cryoboralfs in high altitude forests in Wyoming (Yavitt and Fahey 1986). Evans *et al.* (2000) found total dissolved phosphorus (TDP) concentrations in excess of 800 $\mu g\ L^{-1}$ in groundwater flow beneath ephemeral stream channels in a complex soil landscape in NE Alberta. This latter case reflected the importance of soil organic matter as a source of P. Shaw (1989) reported that mean SRP concentrations varied from 0.06 to 2.1 mg $L^{-1}$ in bottom sediments of six Boreal Plain lakes. A wide range in concentration of dissolved P has been reported in typical boreal streams in western Canada. Munn and Prepas (1986) reported mean TDP in two boreal streams in west central Alberta of just 12 $\mu g\ L^{-1}$ in a region dominated by conifer vegetation, Gray Luvisols and Organic soils (Knapik and Lindsay 1983). At another location on the Boreal Plain dominated by Gray Luvisolic and Organic soils but aspen vegetation, Cooke and Prepas (1998) found annual average dissolved P concentration to vary from 50 to 93 $\mu g\ L^{-1}$ in two forest streams. P movement within landscapes is important but will not be revealed with watershed scale studies if one part of the landscape acts as a sink for P generated elsewhere. Furthermore, changes to the importance of hydrologic flowpaths following forest disturbance (e.g., increased overland flow) influence P mobility.

***Objectives and Hypotheses***

Movement of P at the hillslope scale is needed to understand movement of P in Boreal Plain watersheds. The objective of this study was to investigate P solubility in typical horizons of Gray Luvisolic soil profiles. I hypothesized that:

- Soluble P concentration would be greatest in upper horizons due to weaker and less abundant sorption sites

- P in upper soil horizons would be more soluble than P in lower horizons
- Soluble P would be the most important form of P in subsurface flow
- Organic P would dominate the soil solution

Soil solution composition was determined with saturated paste extracts from soil horizons and from *in situ* samples obtained from groundwater or subsurface flow. P adsorption and desorption characteristics were determined by a batch process and analyzed with the Freundlich and Langmuir equations (Riemsdijk 1996). P speciation modeled with GEOCHEM (Sposito and Mattigod 1979) was plotted on phosphate stability diagrams (Lindsay and Moreno 1960; Lindsay 1979) to infer the probable solid phase controlling P in solution.

### *Materials and Methods*

Field work was carried out in the Boreal Mixedwood Ecoregion (Strong and Leggat 1992) near Lesser Slave Lake, Alberta, Canada. Mean annual precipitation at Lesser Slave Lake for the period 1950 to 1980 was 488 mm of which 35% fell as snow and mean July temperature was 15.6 °C (Environment Canada 2000). Gray Luvisolic soils dominate upland locations and Organic soils dominate low areas in the ecoregion (Strong and Leggat 1992). Vegetation in upland areas was primarily aspen (*Populus tremuloides* Michx.) and white spruce (*Picea glauca* [Moench] Voss). This location was chosen for its fine textured soils, rolling topography and proximity to a set of lakes undergoing studies related to forest harvest and the role of buffer strips in protecting water quality (Prepas *et al.* 2001). A hillslope segment of approximately 50 m length and 0.5 ha area was chosen for a detailed examination of subsurface flow processes and P distribution after forest harvest. The hillslope had convex slope profile and curvature, 13% gradient at midslope, and an easterly aspect. The water table was approximately 5 m below surface at the lower slope position. The adjacent lake lay at the foot of this hillslope, approximately five m lower in elevation and 30 m distant from the nearest instruments. Timber was harvested in January 1997, with a combination of feller-bunchers, skidders and forwarders. Larger slash accumulated near haul roads and remained thereafter to decompose naturally. Other than soil compaction, which was reflected by bulk densities in the upper mineral soil an average of 16% higher than forested sites three years after harvest (Whitson *et al.* 2003), soil horizons were not disturbed by the harvesting activity. No silivicultural treatment was employed on the site except the planting of white spruce seedlings.

To sample the soil solution *in situ*, stainless steel zero-tension flow collectors were installed in June 1998, in the LFH horizon (2.44 m x 0.15 m) and within the upper and lower half of the Ae horizon (2.44 m x 0.05 m) at the lower-midslope position at each of three transects (Figure 25). Flow collectors intercepted saturated subsurface flow associated with these horizons during the spring snowmelt and during rainfall events over the period September 1998 to October 1999. Flow was routed through high-density polyethylene tubing to collection bottles of similar high density polyethylene. Groundwater was sampled with plastic (PVC) wells installed to a depth of 5.2 m below surface at the lower slope position where reaching the groundwater table by hand augering was feasible. The well stem from 4.0 to 5.2 m depth was wrapped in geotextile fabric and screened with sand. The well stem from 0.5 to 4.0 m was surrounded with pure bentonite sealant to prevent groundwater contamination. Wells were installed in the autumn of 1998 and samples were collected approximately twice monthly during 1999. After pH measurement, samples were suctioned through pre-rinsed Millipore HAWP 0.45-µm filters and immediately frozen. *In situ* soil water samples were measured for total phosphorus (TP), TDP, and soluble reactive phosphorus (SRP). Groundwater samples were analyzed for TDP and TP only. TP and TDP samples received a persulfate digestion (Prepas and Rigler 1982) before analysis. A total of 54 SRP, 57 TP, and 66 TDP samples were obtained. SRP, TDP, and TP (after digestion) were determined by a Cary 50-probe UV-Visible Spectrophotometer with the modified ascorbic acid method (Murphy and Riley 1962) and with light absorbance measured at 885 nm.

The second approach to characterizing the soil solution was to obtain soil samples from several slope positions (Figure 25). Sampling was conducted in 4 strata based upon slope position: upper, upper-mid, lower-mid, and lower. Three transects running parallel from the crest to the lower slope position allowed sampling replication within strata (Figure 25). Soil profiles were classified to subgroup level of the Canadian system of Soil Classification (Agriculture Canada Expert Committee on Soil Survey 1987) and landform was identified with reference to parent material characteristics to further classify profiles to the series level. Soil profiles were sampled by horizons to a depth of 1.2 m at upper slope, 2.0 m at upper-midslope, 3.0 m at lower-midslope, and 5.2 m at lower slope positions. The upper and lower half of the Ae horizon was sampled separately at the lower-mid slope position. Soil samples were air dried, ground to pass through a 2-mm sieve, and stored at room temperature until analysis. Analytical methods included particle size analysis by the hydrometer method (Sheldrick and Wang 1993), $H^+$ ion activity in 1:2 soil:solution ratio and 0.01 M $CaCl_2$ with an Accumet 925 pH meter, and electrical conductance (EC) by electrode and 1:2 soil to solution ratio. Select profiles were examined for exchangeable cations by the ammonium acetate method (Hendershot *et al.* 1993), inorganic and organic carbon with a Carlo-Erba Carbon and Nitrogen elemental analyzer, and total P and nitrogen by Technicon Autoanalyzer after Kjeldahl digestion. Available P was determined for LFH and Ae horizons with an extracting solution of 0.03 N $NH_4F/H_2SO_4$, and a 1:5 soil to solution suspension subjected to 10 minutes of shaking.

Soil solution was obtained with saturated paste extracts (Janzen 1993) prepared with de-ionized water and acid-rinsed glassware (3 % HCl). Extracts were suctioned through pre-rinsed 0.45-μm Millipore HAWP filters after routine pH (Accumet 925 metre) and EC measurement (Radiometer CDM 83 Conductivity Meter) and stored at 5° C until analysis. Filtered extracts were analyzed for cations, anions, and dissolved organic carbon (DOC) to provide a dataset that would allow P speciation modeling. Where extract volume was insufficient, samples from the same horizon and slope position were combined. Calcium ($Ca^{2+}$), magnesium ($Mg^{2+}$), sodium ($Na^+$), potassium ($K^+$), iron ($Fe^{3+}$), and manganese ($Mn^{2+}$) were analyzed by atomic absorption spectrometry while soluble Aluminum ($Al^{3+}$)

was determined by atomic absorption spectrometry in conjunction with a graphite furnace (Zhang and Taylor 1989). Sulfate ($SO_4^{2-}$), chloride ($Cl^-$), and fluoride ($F^-$) were determined by ion chromatography (O'Dell *et al.* 1993). DOC was measured by the combustion/infrared technique with a Total Organic Carbon Analyzer (Greenburg *et al.* 1992). Silicon ($Si^{4+}$) was determined by the ammonium molybdate-stannous chloride method (Stainton *et al.* 1977) with absorbance read at 820 nm on a Cary 50-probe UV-Visible Spectrophotometer. Some duplicate samples were analyzed for P by ion chromatograph. The effect of freezing on P concentrations was determined for a subset of samples representing a broad range.

A P sorption study was conducted on horizons from select profiles. A 1:10 soil to solution ratio was employed with 0.01 M $CaCl_2$ as the background electrolyte. After agitation with a wrist-action shaker, samples were centrifuged and solutions decanted and analyzed for P by the ascorbic acid method at 885 nm on a Cary 50-probe UV-Visible Spectrophotometer. Surface excess was assumed to be the difference between the original quantity of P in solution and the amount remaining at equilibrium. All measurements were conducted in triplicate to reduce analytical error. The times required to reach adsorption and desorption equilibrium were determined beforehand for the Ae, BC and Ck1 horizons. Samples were equilibrated in a solution with an initial P concentration of 10 mg $L^{-1}$ for 0, 12, 24, 48 and 72 h. The time required to reach desorption equilibrium was determined with samples previously equilibrated to 10 mg $L^{-1}$ P solution. The original P solution was decanted and replaced by 0.01 M $CaCl_2$ solution and again subjected to shaking for periods of 4, 18, 24, 48 and 72 h and then analyzed as before.

For the adsorption study, 4 g soil and 40 mL 0.01 M $CaCl_2$ solution containing graded amounts of $KH_2PO_4$ were added to 50 mL centrifuge tubes. Initial P concentrations for adsorption were varied from 0 to 100 mg $L^{-1}$ for Bt, BC, Ck1 and Ck3 horizons and from 0.01 to 50 mg $L^{-1}$ for the Ae horizon. Ae horizons were equilibrated for 48 h with a wrist action shaker while Bt, BC, Ck1 and Ck3 horizons were equilibrated for 24 h. Samples were centrifuged and decanted, after which more 0.01 M $CaCl_2$ solution was added. The quantity of solution to add was determined with reference to the volume of solution remaining in the soil to maintain a 10:1 soil to solution ratio. The quantity of P carried forward from the previous adsorption study was determined and accounted for in subsequent calculations. Samples of all horizons were shaken for 18 h to reach desorption equilibrium.

Sorption data (solution concentration at equilibrium in units of mg $L^{-1}$ plotted against surface excess in mg $kg^{-1}$) were analyzed with both Freundlich and Langmuir equations (Sibbesen 1981). The Freundlich equation is a power equation of the form:

$$Y = ax^n \text{ where } 0 < n < 1 \qquad [1]$$

Y, the surface excess, is defined as the weight of P sorbed per unit weight of soil, x is the P solution concentration at equilibrium, and a and n are constants determined for each soil sample (Sposito 1984). The log-transformed equation was used to solve for these constants:

$$\text{Log } Y = \log a + n \log x \qquad [2]$$

The constant "a" is thus equivalent to surface excess at a solution concentration of 1 mg $L^{-1}$ P. The Langmuir equation has long been used to describe P sorption in soil (Parfitt 1978; Ryden and Pratt 1980) and is a hyperbolic equation of the form:

$$x/m = kCb/1+kC \quad [3]$$

where x/m is the surface excess, C is solution concentration, k is related to P binding strength and b is the adsorption maxima. The values of coefficients are determined with the linear version of the Lanmuir equation:

$$C/(x/m) = 1/Kb + C/b \quad [4]$$

Despite its widespread use, the Langmuir equation oversimplifies the adsorption process and may have limited physical meaning (Harter and Smith 1981; Sposito 1984).

The adsorption isotherms were also analyzed to determine the value of the equilibrium P concentration and the amount of surface P already present, or native P (Figure 26). Both equilibrium P concentration and native P were estimated by extrapolating the linear portion of the isotherm to a theoretical solution concentration of zero. Native P content was added to the value of the surface excess prior to determination of sorption coefficients (Bridgham *et al.* 2001) and resulted in improved fit of Langmuir and Freundlich equations. Differences in sorption characteristics among horizons were determined by a one-way ANOVA followed by a Tukey test. Differences were significant at a probability of less than or equal to 0.05.

The activity of P complexes in the soil solution was modeled with GEOCHEM (Sposito and Mattigod 1979). Ion activity was estimated from mean solute concentrations of each horizon with the Davies equation. Total carbonate in solution was estimated from the relationships among $H_2CO_3$, $HCO_3^-$ and $CO_3^{2-}$ and dissolved carbon dioxide in equilibrium with an atmospheric $CO_2$ of partial pressure of 30.3 Pa (Lindsay 1979). Ionic strength (I) was estimated from specific conductance (EC) of saturated paste extracts with the equation

$I=0.013EC$ where EC is in units of $dS\ m^{-1}$.

Stability diagrams for the most probable solid phase P minerals were constructed for the LFH, upper and lower Ae, Bt, BC, and Ck1 horizons with reference to thermodynamic stability constants reported by Lindsay (1979). Potential solid phases were assumed to possess unit activity. Estimated orthophosphate activity ($HPO_4^{2-}$ or $H_2PO_4^-$, depending on pH) was then plotted in relation to these solid phases. Key P minerals were assumed to include brushite ($CaHPO_4.2H_2O$), monetite ($CaHPO_4$), octacalcium phosphate ($Ca_4H(PO_4)_3.2.5H_2O$), hydroxylapatite ($Ca_5(PO_4)_3OH$), fluoroapatite ($Ca_5(PO_4)_3F$), strengite ($FePO_4.2H_2O$), and variscite ($AlPO_4.2H_2O$). The presence of apatite was expected given its stability under the high pH of parent materials at the site. Activities of Ca, Fe, and Al were plotted on stability diagrams for their respective potential solid phases to infer which of these minerals controlled their solubility. Anions that were likely in equilibrium with phosphates ($F^-$, $Si^{4+}$) were estimated from their calculated activity in solution. Ca in the Ck horizon was assumed to be controlled by calcite and $CO_2$ partial pressure, which was assumed to be 0.0003 atm (30.3 Pa) (atmospheric level) in all horizons.

## *Results*

### *Profile Characteristics*

Soil profiles at all slope positions of this morainal hillslope were classified as Orthic Gray Luvisols of the Grandin series. The typical horizon sequence was an LFH-Ae-Bt-BC-Ck and there were pronounced differences in soil properties with depth (Table 10). The Ae horizon had lower clay content than the underlying Bt, BC and Ck1 horizons, which were all dominated by the clay fraction. Calcium carbonate content of Ck1 horizons averaged 1.4 % and was encountered at an average depth of 178 cm. pH varied with horizon, with lowest pH in the Bt and BC horizon and highest in the unweathered Ck horizon. The LFH and Ae horizons were slightly acidic. The ground water zone (Ck3) was similar to the Ck1 and Ck2 horizons except for duller colors and CL texture. EC was less than 0.2 dS $m^{-1}$ in all horizons. The order of dominance of exchangeable cations in the solumn was: calcium>magnesium>>potassium>sodium.

Soil nutrient properties also varied strongly with depth (Table 11). OC concentration of the LFH horizon was more than 50 times higher than other horizons. Total N concentration was nearly 20 times higher and total P concentration was 2 to 3 times higher in the LFH than other horizons. Upper and lower Ae horizon increments were similar for texture and pH (Table 12). Exchangeable cation concentration, particularly $Ca^{2+}$ and OC were higher (1.4 and 3 times respectively) in the upper increment than the lower. Soil properties on this hillslope were similar to those observed for other Gray Luvisolic profiles studied on the Boreal Plain (Pawluk 1961; Pawluk and Bayrock 1969; Abder-Ruhman 1980; Spiers 1982; Howitt and Pawluk 1985a; Howitt and Pawluk 1985b). Clay transport from the Ae to the Bt is the primary pedogenic process. Divalent base cations dominate the exchange complex even though the solumn reaction was slightly acidic. Higher exchangeable cation concentrations in the upper Ae horizon reflect the important contribution of organic carbon to cation exchange capacity. Available P concentrations were approximately 2 times higher in the LFH than the Ae on the two occasions measured (Table 13).

*In Situ Samples*

There was considerable scatter in the values of both SRP and TDP in all horizons except the Ck3 (Figure 28 and 29). Average SRP in the LFH was 4.0 mg $L^{-1}$, with approximately half of the values between 2 and 6 mg $L^{-1}$ (Table 14). Average SRP in the upper Ae horizon was 4.1 mg $L^{-1}$, with approximately half the values between 3 and 5 mg $L^{-1}$. SRP was somewhat less in the lower Ae horizon, averaging 1.2 mg $L^{-1}$, and with more than 75% of values less than 2 mg $L^{-1}$ (Table 14). Trends were similar for total dissolved P. Mean values of LFH and upper Ae horizons were highest (4.1 and 4.0 mg $L^{-1}$ respectively), intermediate for lower Ae horizons (1.3 mg $L^{-1}$) and least in the Ck3 horizon (0.01 mg $L^{-1}$). Variation in TDP within horizons was almost identical to SRP. The ratio of TDP to TP was 0.84 in the LFH, 0.90 in the upper Ae, 0.88 in the lower Ae, and 0.76 in the Ck3, indicating that more than 75% of P occurred in the dissolved phase. The ratio of SRP to TDP was 0.95 in the LFH, 0.99 in the upper Ae, and 0.90 in the lower Ae.

SRP of *in situ* samples decreased over time in the LFH horizon within the April to May and August periods when precipitation was greatest (Figure 28a). Both upper and lower Ae horizon showed a similar downward trend in the April to May period (Figure 28b). TDP trends were similar, except that TDP concentrations in the groundwater zone remained stable (Figure 29a-b).

*Soil Extract Solution Characteristics*

EC was greatest in LFH and Ck horizons (Table 15). The pH was above 8 for the Ck horizons, and generally neutral to slightly acidic for Ae and LFH horizons, and slightly acidic in Bt and BC extracts. DOC concentration decreased with depth, averaging 1513 mg $L^{-1}$ in the LFH horizon extracts and reaching a minimum of 38 mg $L^{-1}$ in extracts from the groundwater zone. Calcium was the dominant cation, followed by magnesium and potassium. Highest absolute amount of cations were found in the LFH and Ae horizons. Al concentrations were marginally higher in LFH and Ae horizons than for the Ck horizons. The quantity of cations (milliequivalent basis) was similar to the amount of measured anions (sulfate, chloride, fluoride, P), even though organic acids and carbonates were not included in the charge balance calculations. SRP was most concentrated in the LFH horizon where it was the dominant measured anion, and decreased sharply in the other horizons (Table 16). Concentrations in the Ae horizon were higher than in the subsoil and in the Ck horizons. The relationship between log transformed DOC and SRP was very strong, with an $r^2$ of 0.93 (Figure 30). Freezing had no measurable effect on SRP, with the relationship between frozen and non-frozen samples possessing a slope of 1.01 and an $r^2$ of 1.0. Both ion chromatrograph and standard colorimetric techniques yielded equivalent P values (Figure 31). The latter finding suggests that orthophosphate may be the dominant form of P in solution, and also suggest that the spectrophotometer readings were not influenced by the relatively high DOC content.

*Sorption Characteristics*

Sorption characteristics differed among soil horizons. The Freundlich equation (Table 17) tended to fit the experimental results better than the Langmuir equation (Table 18). Mean values of $r^2$ for the Freundlich equation varied from 0.97 to 1.00 during adsorption and from 0.98 to 0.99 for desorption (Table 17). Mean value of $r^2$ for the Langmuir equation ranged from 0.95 to 1.0 for adsorption and from 0.91 to 0.99 during desorption (Table 18). The Freundlich intercept 'a' value, an index of P retention at an equilibrium solution concentration of 1 mg $L^{-1}$, was lower in the lower Ae horizon than the Bt, BC and Ck horizons ($P$<0.05). Values were approximately 1.5 times higher during desorption, and ranking among horizons remained similar. Adsorption maxima were lower in the lower Ae than the other horizons ($P \leq 0.01$). The Langmuir coefficient for binding strength during adsorption was higher in the Bt horizon ($P$<0.03). Desorption values were highest in the BC, lowest in the upper Ae, and intermediate in the remaining horizons ($P$<0.001). Desorption maxima were similar among all soil horizons ($P$=0.28). Trends between adsorption and desorption maxima varied by horizon. Desorption maxima standard deviation were particularly high in the upper and lower Ae horizons. $P_{eq}$ concentration was higher in the upper Ae horizon than other horizons during adsorption and desorption ($P \leq 0.01$). Equilibrium P concentrations tended to be lower for desorption. Standard deviation of $P_{eq}$ was more than an order of magnitude higher in the upper and lower Ae than other horizons.

*Forms of P in Solution Predicted by GEOCHEM*

The inference that most dissolved P was orthophosphate was supported by chemical speciation modeling (Figure 32). GEOCHEM predicted orthophospohate to vary from 71% in the LFH, 96% in the upper and lower Ae, 98% in the Bt, 99% in the BC, and 62% in the Ck1 of the total (data not shown). Some 18% of measured P was predicted to precipitate as Ca-phosphate in the LFH, and 38% of P existed as Ca- and Mg-phosphate complexes in the Ck1 horizon soil solution. The activity of Al- and Fe-phosphate complexes decreased sharply between the BC and Ck1 horizons.

The activity of P in solution relative to potential solid phase phosphates is shown in Figure 33 a-d. The activity of $Si^{4+}$, $Al^{3+}$, $Fe^{3+}$, and $Ca^{2+}$ involved in the formation of these solid phase phosphates were estimated from measured concentrations and plotted on stability diagrams to infer the solid phase controlling their respective equilibrium (Table 19). $Fe^{3+}$ appeared to be in equilibrium with amorphous iron in all horizons except the LFH, which appeared to be in equilibrium with maghemite (data not shown). $Al^{3+}$ appeared to be in equilibrium with gibbsite in all horizons except the Ck1, where its activity was bounded by the kaolinite and gibbsite isotherms (data not shown). $H_2PO4^-$ activity of the LFH intersected the strengite, brushite, Octacalcium phosphate, and monetite isotherms, indicating an equilibrium between several potential Fe- and Ca-phosphates. $H_2PO4^-$ activity of the upper Ae was in equilibrium with fluorapatite, strengite and variscite, while $H_2PO_4^-$ activity in the lower Ae was in equilibrium with fluorapatite only. $H_2PO_4^-$ in Bt and BC (not shown) horizons plotted below the fluorapatite-variscite-strengite isotherms. In the Ck1 horizon, $HPO_4^{2-}$ activity resided above the hydroxylapatite isotherm, indicating oversaturation with respect to these minerals (data not shown).

### *Discussion*

Very high P concentrations from extracts and *in situ* samples from LFH horizon and the general steep decline in concentration with depth support the hypothesis of very high solution P in upper horizons. The high TDP to TP ratio for *in situ* samples supports the hypothesis of the importance of the soluble phase in P transport. Saturated paste extracts averaged approximately three orders of magnitude greater in LFH horizons than in Ae horizons, and four orders of magnitude greater than subsoil, parent material and the groundwater zone. $P_{eq}$ concentrations in the upper Ae horizon were 55 to 70 fold greater than those in the groundwater zone. Reduction in solution P concentration with depth reflects changes to sorption characteristics and to the forms of solid phase P in equilibrium with solution P.

In general, dissolved P has been considered immobile because of typically low soil solution concentrations (Peaslee and Phillips 1981). In this study, SRP and TDP concentrations observed for upper soil horizons one to two orders of magnitude higher than reported for other climatic and physiographic zones. P concentration in leaf litter leachate from spruce plantations on acidic Podzolic soils in Wales after harvest ranged from near 0 to 0.6 mg $L^{-1}$ (Stevens *et al.* 1995). Cortina *et al.* (1995) observed SRP in pine plantation forest floor leachate in Spain to vary from 0.02 to 2.50 mg $L^{-1}$, with highest values in summer. TDP concentrations in runoff from Birch forests in Minnesota averaged 0.33 mg $L^{-1}$ from litter layers and 0.07 mg $L^{-1}$ in Ae horizon interflow (Timmons *et al.* 1977). High P values in both *in situ* and soil extracts may reflect reduced leaching due to dry weather conditions of the study period, which produced only 50% of the 1961 to 1990 average precipitation, or they may reflect acclerated mineralization following logging. I found that most roots were restricted to the upper 20 cm of the soil, corresponding to the LFH and Ae horizons. Similar findings are reported by others (Strong and La Roi 1983; Huang and Schoenau 1997). High P concentrations within the LFH horizon, while beneficial to plants, also suggest that water movement in this part of the profile could lead to significant downslope transport.

While the strong correlation between SRP and DOC concentrations supported the hypothesis of organic P dominance in solution, other explanations are also suggested by the data. The high ratio of SRP to TDP suggests that a majority of dissolved P in the *in situ* samples was orthophosphate. However, SRP concentration can include forms of P in addition to orthophosphate (Dick and Tabatabai 1978). Nor does the apparent existence of P in the dissolved phase mean that the species measured as SRP are necessarily ionic; P may be colloidal (Haygarth *et al.* 1997; Sinaj *et al.* 1998). However, the 1:1 relationship observed between paired samples in this study examined by both the colorimetric technique and ion chromatography provides strong evidence that SRP was orthophosphate. Furthermore, the prediction of orthophosphate dominance by GEOCHEM lends credence to empirical findings of inorganic P dominance. Nevertheless, the high ratio of SRP to TDP in both *in situ* samples and soil horizon extracts is at odds with the literature for forested ecosystems. Smith *et al.* (1998) reported that SRP made up no more than 65% of total P in leachate under black spruce forests in Quebec, Canada. Timmons *et al.* (1977) found that SRP made up just 20% of total dissolved P in a birch forest leachate in Minnesota. Yavitt and Fahey (1986) found that SRP made up less than 5% of soluble P in litter layer leachate in coniferous forests. The high proportion of SRP ($\geq$ 90%) in the dissolved P fraction of sampled upper horizons from my study might have resulted from accelerated mineralization following harvest, or due to P release that occurred during soil handling and drying. DOC produced in the LFH horizon probably has a major influence on P mobility in Gray Luvisols, if not in producing organic P, then by initiating reactions that effectively make P more soluble. DOC will enhance conditions for P-metal-organic complex transport (Schnitzer 1969). DOC can displace P from sorption sites and complex metal cations that would otherwise complex or precipitate P (Pohlman and McColl 1988; Fox 1993).

Higher SRP in soil extracts than *in situ* samples partially resulted from differences in sampling conditions between the two approaches. *In situ* samples were obtained at lower temperature, often near $0^{\circ}$ C, and the period of equilibration between soil and water was unknown and likely highly variable reflecting differences in rainfall intensity or snowmelt rate, antecedent soil water content, and hydraulic conductivity at the individual sites. Relative to soil extracts, *in situ* samples appeared to be P-undersaturated in the LFH, P-oversaturated in the upper and lower Ae, and near equilibrium in the groundwater zone. Higher dissolved P concentrations during snowmelt than summer may also reflect seasonal differences in biological uptake of P. Soil extracts may also reflect P mineralization by microorganisms during the drying, handling, and extraction process. Alternatively, differences may reflect kinetic control of P solubility, with *in situ* samples tending to represent non-equilibrium conditions.

Differences in sorption magnitude and bonding energy among soil horizons provide some support for the hypothesis that sorption reactions are more important in subsoils and the groundwater zone. Changes in texture, pH and mineralogy with depth mean that mechanisms of adsorption and desorption are likely to change with depth. However, these mechanisms are not revealed by empirical formulations such as the Freundlich and

Langmuir equations. It is likely that the reduction in clay surface area in Ae horizons is balanced by increases in surface area of minerals that have much higher P affinity and/or site density, such as amorphous iron and aluminum oxides. Differences in desorption characteristics among horizons were less than for adsorption. Desorption isotherms were usually shifted to the left, reflecting the irreversibility of P adsorption. In general, Ae horizons had lowest capacity to sorb P and weakest bonding energy. The relatively high standard deviation of available P contents of the Ae horizon were probably a reflection of unequal inputs of P from litter layers and subsequent adsorption by mineral soil.

As an alternative to sorption reactions, solution P may be in equilibrium with potential solid phase forms. The assumption of equilibrium that underlies use of stability diagram is reasonable for this forested region, which has existed in a natural cycle of forest succession interrupted by occasional wildfire, and most recently by logging, with most P inherited rather than recently incorporated. The equilibrium between $Fe^{3+}$ and maghemite for the LFH horizon is probably a reflection of past forest fires and conversion of other iron oxides to this form (Ketterings *et al.* 2000). The change of potential solid phases phosphates with depth are consistent with results of soil weathering and pedogenesis, with secondary Ca- and Fe-phosphates in the LFH replacing primary Ca-P (fluoroapatite) as weathering proceeds. The apparent control of fluorapatite on P solubility in the Ae horizon is reasonable given the relatively limited chemical weathering that leaves some primary P still present (Schoenau *et al.* 1989). Apparent nonequilibrium of P with either fluorapatite or variscite in the Bt and BC horizons is best explained by P adsorption (Sposito 1984). Finally, the nonequilibrium between postulated solid-phase phosphates and orthophosphate activity in the Ck1 horizon is not understood. The lack of weathering in this part of the profile may have meant that soil minerals no longer present in the solum, such as carbonates, control P solubility in the parent materials.

My results have implications for the effectiveness of vertical and lateral subsurface P transport on Boreal Plain hillslopes with Gray Luvisols. Significant P can reach the subsoil if soil water is traveling rapidly along macropores (Jensen *et al.* 1999). Lateral flow at the LFH/Ae interface would probably have the highest P concentrations. Such a flow process is likely during snowmelt when soil is frozen soil or when rainfall intensity exceeds infiltration capacity of the Ae. The potential of the latter increases with both logging-induced soil compaction (Whitson *et al.* 2003) and post-harvest site preparation activities. Interflow occurring at the Ae-Bt horizon interface, the most probable location of subsurface flow (Whipkey and Kirkby 1978) would have lower P concentration than flow at the LFH-Ae interface, but still greater than the subsoil or groundwater zone. The low P concentration in the groundwater zone will mean that much more groundwater flow will be needed to transport as much P moved by shallow subsurface flow.

***Conclusions***

Strong vertical P stratification in Orthic Gray Luvisolic soils reflects differences in sorption and P solubility properties with depth. Soil weathering and soil biota have converted P into more soluble forms in upper horizons such that P concentrations are more than 5000 times higher in the LFH than lower in the profile. The increased solubility of P in the LFH horizon is explained by the development of secondary and organic P of greater solubility than primary P minerals present at greater depths. Higher clay or carbonate contents in subsoils or parent materials will lead to stronger adsorption reactions. Given the pedogenic environment associated with Gray Luvisols, these soils may be sources of P for locations lower in the landscape.

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Table 9: Definitions of terms used

| | |
|---|---|
| a | Intercept of the Freundlich equation |
| b | Adsorption maxima of the Langmuir equation |
| K | Coefficient of binding strength for the Langmuir equation |
| n | Exponent of the Freundlich equation |
| Organic P | Organic molecules with phosphorus present in functional group (e.g., typically inositol P, nucleic acids, phospholipids from microorganisms) |
| orthophosphate | $H_3PO_4$; phosphoric acid. Typically present in soil as either $H_2PO_4^-$ or $HPO_4^{2-}$ |
| $P_{eq}$ | SRP concentration measured in a soil solution at equilibrium |
| soluble P | P remaining in solution after filtering with 0.45-µm filters |
| solum | Soil horizons above the unweathered parent material |
| SRP | Soluble reactive phosphorus: that component of total dissolved P measured with the ascorbic acid method prior to digestion |
| subsoil | Bt and BC horizons |
| TDP | Total dissolved phosphorus: the quantity of P in solution found by the ascorbic acid method after persulfate digestion of filtered samples |
| Total P | Quantity of P found by ascorbic acid determination of an unfiltered, digested sample |

Table 10: Properties of hillslope soils: mean and (*SD*); nd = no data

| Horizon | | Clay | Silt | Sand | pH ($CaCl_2$) | EC dS $m^{-1}$ | Exchangeable Cations (meq 100 $g^{-1}$) | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| Label | Thickness (cm) | | % | | | | Ca | Mg | K | Na |
| LFH | 11 (3) | nd | nd | nd | 5.6 (0.3) | nd | 64.1 (6.4) | 10.8 (1.8) | 2.8 (0.3) | trace (0.02) |
| Ae | 16 (4) | 15 (4) | 64 (7) | 21 (5) | 5.2 (0.7) | 0.07 (0.02) | 7.1 (3.3) | 1.6 (0.4) | 0.1 (0.2) | trace (0.04) |
| Bt | 44 (6) | 44 (6) | 32 (8) | 24 (4) | 4.8 (0.5) | 0.05 (0.00) | 18.3 (0.4) | 9.8 (1.0) | 0.05 (0.07) | 0.03 (0.09) |
| BC | 118 (13) | 44 (3) | 32 (4) | 24 (4) | 4.7 (0.4) | 0.05 (0.01) | 18.8 (0.9) | 11.4 (0.6) | 0.01 (0.07) | 0.04 (0.02) |
| Ck1 | 87 (12) | 42 (2) | 31 (2) | 27 (4) | 7.5 (0.1) | 0.18 (0.01) | nd | nd | nd | nd |
| Ck2 | 138 (49) | 36 (0) | 32 (2) | 32 (1) | 7.6 (0.1) | nd | nd | nd | nd | nd |
| Ck3 | nd | 31 (5) | 31 (1) | 38 (6) | 7.5 (0.1) | nd | nd | nd | nd | nd |

Table 11: Chemical properties of Gray Luvisolic soils on hillslope: mean and (*SD*)

| Horizon | Organic Carbon | Total Nitrogen | Total Phosphorus |
|---|---|---|---|
| | % | | $mg\ kg^{-1}$ |
| LFH | 40.4 | 1.8 | 1350 |
| | (5.4) | (0.2) | (67) |
| Ae | 0.8 | 0.1 | 610 |
| | (0.4) | (0.02) | (108) |
| Bt | 0.6 | 0.1 | 420 |
| | (0.03) | (0.02) | (28) |
| BC | 0.6 | 0.1 | 470 |
| | (0.03) | | |
| Ck1 | 0.5 | 0.09 | 588 |
| | (0.07) | (0.01) | (39) |

Table 12: Properties of Ae horizon: mean and (*SD*)

| Horizon | | Clay | Silt | Sand | pH in $CaCl_2$ | Exchangeable Cations (meq $100g^{-1}$) | | | | Organic C |
|---|---|---|---|---|---|---|---|---|---|---|
| Label | Thickness (cm) | | % | | | Ca | Mg | K | Na | % |
| Upper | 8 | 12 | 62 | 26 | 5.2 | 8.6 | 1.7 | 0.4 | 0.00 | 1.7 |
| Ae | (1) | (1) | (3) | (3) | (0.2) | (1.6) | (0.08) | (0.3) | (0.03) | (0.2) |
| Lower | 9 | 14 | 63 | 23 | 5.1 | 5.9 | 1.5 | trace | trace | 0.6 |
| Ae | (1) | (1) | (5) | (5) | (0.2) | (1.9) | (0.2) | (0.1) | (0.03) | (0.1) |

Table 13: Extractable P (mg $kg^{-1}$)

| Horizon | August 1998 | | October 1998 | |
|---|---|---|---|---|
| | Mean | *SD* | Mean | *SD* |
| LFH | 85.9 | 15.5 | 79.4 | 12.1 |
| Ae | 47.9 | 21.4 | 39.6 | 25.4 |

Table 14: Properties of *in situ* soil water samples: mean and (*SD*); nd=no data

| Horizon | pH | SRP | TDP | TP | TDP $TP^{-1}$ | SRP $TDP^{-1}$ |
|---|---|---|---|---|---|---|
| | | mg $L^{-1}$ | | | | |
| LFH | 7.0 | 4.0 | 4.1 | 5.4 | 0.84 | 0.95 |
| | (0.6) | (2.9) | (2.7) | (3.5) | (0.11) | (0.11) |
| Upper Ae | 6.1 | 4.1 | 4.0 | 4.0 | 0.90 | 0.99 |
| | (0.5) | (1.6) | (1.3) | (2.2) | (0.11) | (0.10) |
| Lower Ae | 6.1 | 1.2 | 1.5 | 2.0 | 0.88 | 0.90 |
| | (0.8) | (1.2) | (1.3) | (1.8) | (0.11) | (0.08) |
| Ck3 | 7.5 | nd | 0.01 | 0.02 | 0.76 | nd |
| | (0.2) | | (0.004) | (0.01) | (0.18) | |

Table 15: Mean (SD) values of soluble cations, anions, pH and EC by horizon as determined with saturated paste extraction (nd = no data)

| Horizon | Ca | Mg | K | Na | $SO_4$ | Cl | F | P | Fe | Mn | Al | DOC | pH | EC |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | (mmol $L^{-1}$) | | | | | | | µmol $L^{-1}$ | | | | mg $L^{-1}$ | | $dSm^{-1}$ |
| LFH | 4.4 | 1.5 | 3.1 | 0.2 | 0.5 | 0.8 | nd | 2074 | 4.8 | 72.1 | 15.3 | 1513 | 6.0 | 0.99 |
| | (0.7) | (0.02) | (0.5) | (0.05) | (0.2) | (0.3) | | (457) | (2.2) | (11.4) | (4.2) | (282) | (0.4) | (0.15) |
| Ae | 1.7 | 0.7 | 0.3 | 0.4 | 0.2 | 0.2 | nd | 7.5 | 7.1 | 39.5 | 14.7 | 311 | 6.9 | 0.42 |
| | (0.6) | (0.08) | (0.1) | (0.05) | (0.05) | (0.05) | | (5.8) | (4.5) | | (5.4) | (42) | (0.5) | (0.13) |
| Upper | 1.4 | 0.4 | 0.4 | 0.2 | 0.4 | 0.9 | nd | 14.9 | 8.8 | 17.8 | 21.5 | 368 | 5.5 | 0.37 |
| Ae | (0.4) | (0.07) | (0.1) | (0.05) | | (0.9) | | | (7.1) | (4.2) | | | (0.2) | (0.09) |
| Lower | 0.8 | 0.2 | 0.2 | 0.2 | nd | nd | nd | 5.4 | 11.6 | 5.6 | 25.5 | 183 | 6.0 | 0.19 |
| Ae | (0.1) | (0.02) | (0.03) | (0.03) | | | | | (3.7) | (2.7) | | | (0.4) | (0.03) |
| Bt | 0.3 | 0.2 | 0.08 | 0.3 | 0.1 | 0.2 | nd | 1.5 | 9.4 | 0.6 | 13.8 | 192 | 5.7 | 0.11 |
| | (0.08) | (0.05) | (0.02) | (0.03) | (0.01) | (0.03) | | (1.5) | (2.7) | | (4.0) | (38) | (0.9) | (0.02) |
| BC | 0.2 | 0.2 | 0.05 | 0.4 | 0.2 | 0.2 | nd | 1.3 | 16.0 | 0.6 | 6.2 | 106 | 5.7 | 0.10 |
| | (0.07) | (0.04) | (0.01) | (0.03) | | (0.04) | | (0.2) | (23.6) | | (5.6) | (28) | (0.8) | (0.03) |
| Ck1 | 1.0 | 0.5 | 0.1 | 0.5 | 0.2 | 0.4 | 0.1 | 0.3 | 0.6 | 0.3 | 7.1 | 59 | 8.3 | 0.29 |
| | (0.2) | (0.07) | (0.01) | (0.03) | | (0.01) | | (0.1) | (0.9) | (0.2) | (3.1) | (1) | (0.3) | (0.03) |
| Ck2 | 1.0 | nd | 0.2 | Nd | nd | nd | 0.08 | 0.3 | nd | nd | 10.0 | 47 | 8.6 | 0.33 |
| Ck3 | Nd | nd | nd | Nd | nd | nd | 0.06 | 0.4 | nd | nd | 5.8 | 38 | 8.5 | 0.48 |

Table 16: SRP in saturated pastes

| Horizon | | SRP (mg $L^{-1}$) | |
|---|---|---|---|
| | | Mean | SD |
| LFH | | 64.2 | 14.1 |
| Ae | | 0.2 | 0.2 |
| | Upper Ae | 0.5 | § |
| | Lower Ae | 0.2 | § |
| Bt | | 0.05 | 0.05 |
| BC | | 0.04 | 0.008 |
| Ck1 | | 0.008 | 0.002 |
| Ck2 | | 0.009 | § |
| Ck3 | | 0.01 | § |

§ composite sample from single slope position

Table 17: Sorption coefficients for Freundlich equation: mean and (*SD*)

| Horizon | a | | n | | $r^2$ | |
|---|---|---|---|---|---|---|
| | Ads | Des | Ads | Des | Ads | Des |
| Upper Ae | 77 ab | 108 ab | 0.42 a | 0.50 | 0.97 | 0.98 |
| | (22) | (30) | (0.04) | (0.07) | | |
| Lower Ae | 57 a | 82 a | 0.39 a | 0.47 | 0.98 | 0.99 |
| | (9) | (13) | (0.02) | (0.04) | | |
| Bt | 98 b | 143 bc | 0.43 a | 0.45 | 0.99 | 0.99 |
| | (19) | (17) | (0.03) | (0.02) | | |
| BC | 115 b | 158 c | 0.32 a | 0.34 | 0.99 | 0.99 |
| | (4) | (5) | (0.004) | (0.002) | | |
| Ck1 | 100 b | 140 bc | 0.35 a | 0.42 | 0.99 | 0.99 |
| | (13) | (23) | (0.02) | (0.03) | | |
| Ck3 | 97 b | 141 bc | 0.33 a | 0.39 | 0.99 | 0.99 |
| | (9) | (9) | (0.02) | (0.02) | | |

Table 18: Sorption coefficients for Langmuir equation and phosphorus equilibrium concentrations: mean and (*SD*)

| Horizon | Langmuir | | | | | | Peq | |
|---|---|---|---|---|---|---|---|---|
| | b | | k | | $r^2$ | | $\mu g\ L^{-1}$ | |
| | Ads | Des | Ads | Des | Ads | Des | Ads | Des |
| Upper Ae | 299 ac | 276 a | 0.48 a | 0.94 ab | 0.99 | 0.99 | 164 ab | 141 a |
| | (23) | (153) | (0.24) | (0.49) | | | (55) | (46) |
| Lower Ae | 220 b | 255 a | 0.49 a | 1.32 abd | 0.95 | 0.91 | 53 bc | 47 b |
| | (15) | (147) | (0.28) | (0.85) | | | (56) | (42) |
| Bt | 313 ac | 427 a | 2.69 b | 1.78 abd | 1.0 | 0.95 | 4 c | 4 b |
| | (17) | (21) | (1.27) | (0.38) | | | (3) | (5) |
| BC | 408 d | 346 a | 1.03 a | 4.21 cd | 0.99 | 0.99 | 3 c | 3 b |
| | (10) | (9) | (0.09) | (0.26) | | | (0.9) | (0.2) |
| Ck1 | 417 d | 360 a | 0.78 a | 1.90 abd | 0.98 | 0.98 | 3 c | 2 b |
| | (16) | (43) | (0.18) | (0.66) | | | (0.5) | (1.4) |
| Ck3 | 371 cd | 328 a | 0.83 a | 2.68 bcd | 0.97 | 0.98 | 3 c | 2 b |
| | (16) | (15) | (0.16) | (0.60) | | | (0.4) | (0.4) |

Table 19: Solid phases or solution activity used for constructing P stability isotherms; nd=no data

| Species | Soil Horizon | | | | | |
|---|---|---|---|---|---|---|
| | LFH | Upper Ae | Lower Ae | Bt | BC | Ck1 |
| $Ca^{2+}$ | measured values | | | | | $CaCO_3$ |
| $Fe^{3+}$ | Maghemite | Amorphous Fe | Amorphous Fe | Amorphous Fe | Amorphous Fe | Amorphous Fe |
| $Al^{3+}$ | Gibbsite | Gibbsite | Gibbsite | Gibbsite | Gibbsite | Kaolinite + Quartz |
| $CO_2^o$ | Atmospheric | | | | | |
| $Si^{4+}$ | nd | nd | nd | nd | nd | -3.66 M |
| $F^-$ | $CaF_2$ | | | | | |

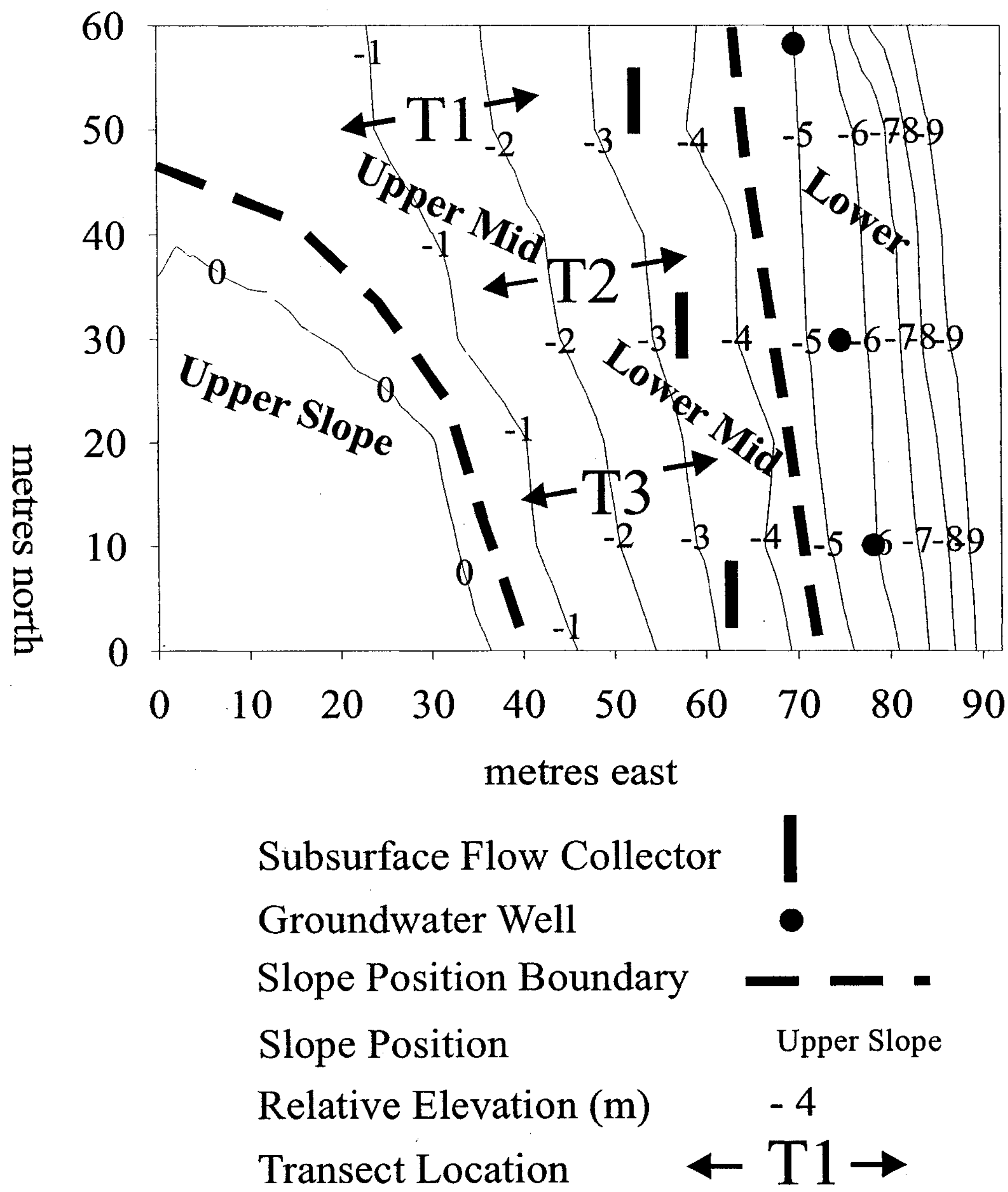


Figure 25: Diagram of hillslope showing slope positions

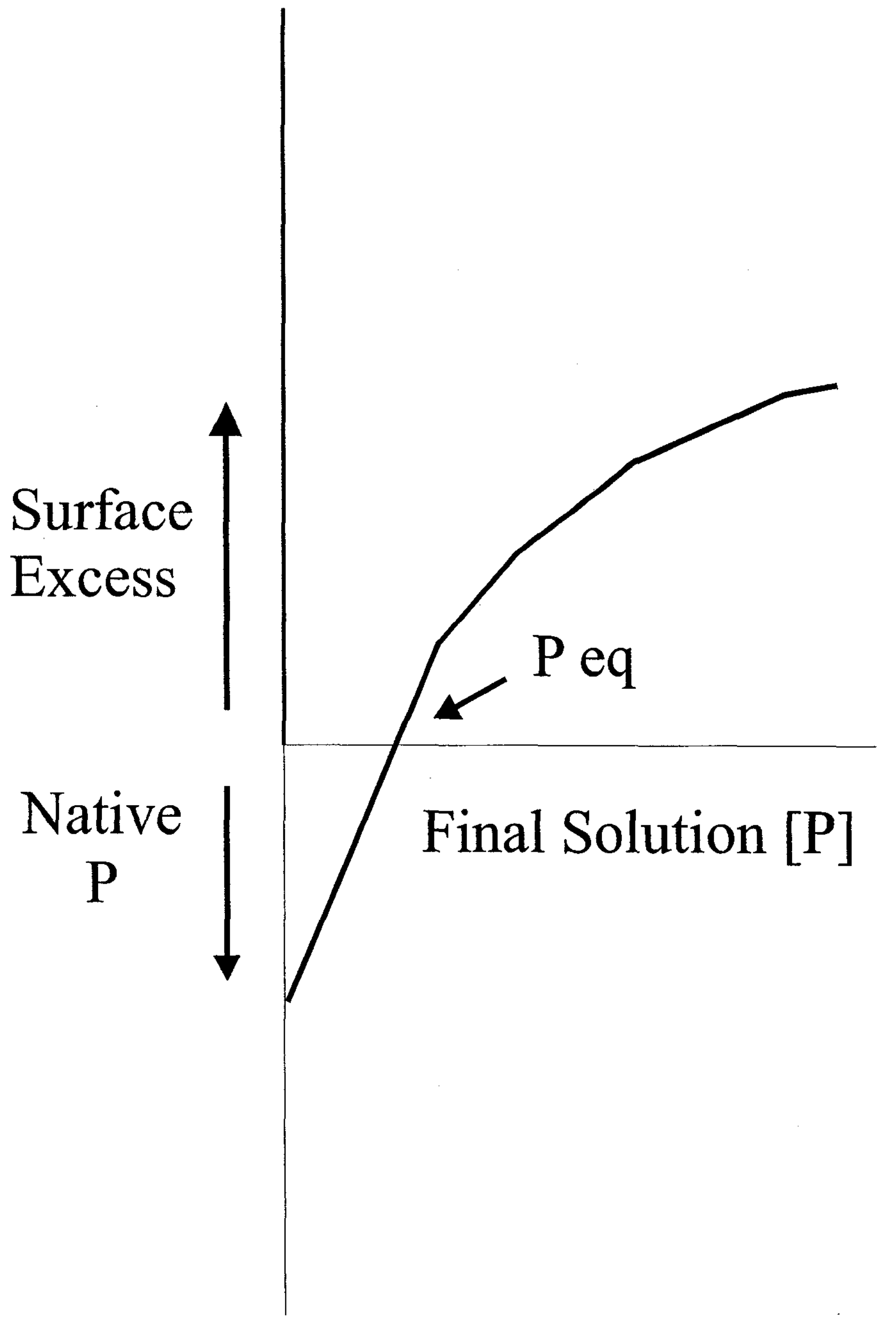


Figure 26: Native P and equilibrium P concentration ($P_{eq}$) determination

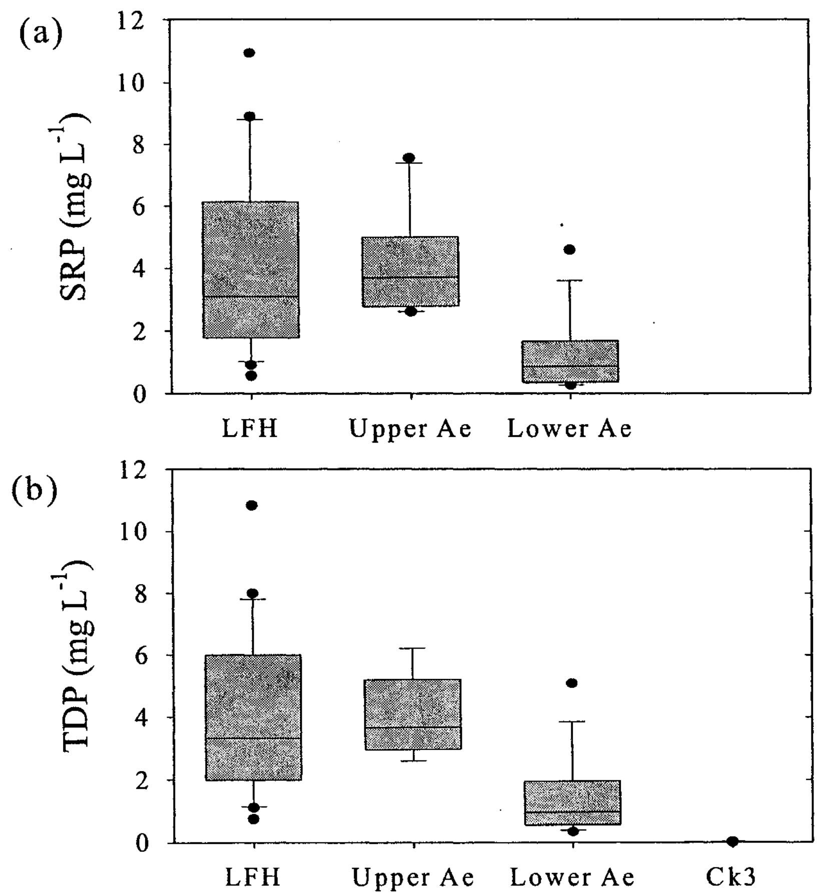


Figure 27: P concentration of *in situ* soil water samples obtained from four horizons, the LFH, upper Ae, and Ck3 (box midline = median, lower and upper box boundaries represent 25$^{th}$ and 75$^{th}$ percentiles, lower and upper whisker caps represent 10$^{th}$ and 90$^{th}$ percentiles, while dots represent outliers) a) soluble reactive phosphorus b) total dissolved phosphorus (Ck3 median = 0.01 mg L$^{-1}$)

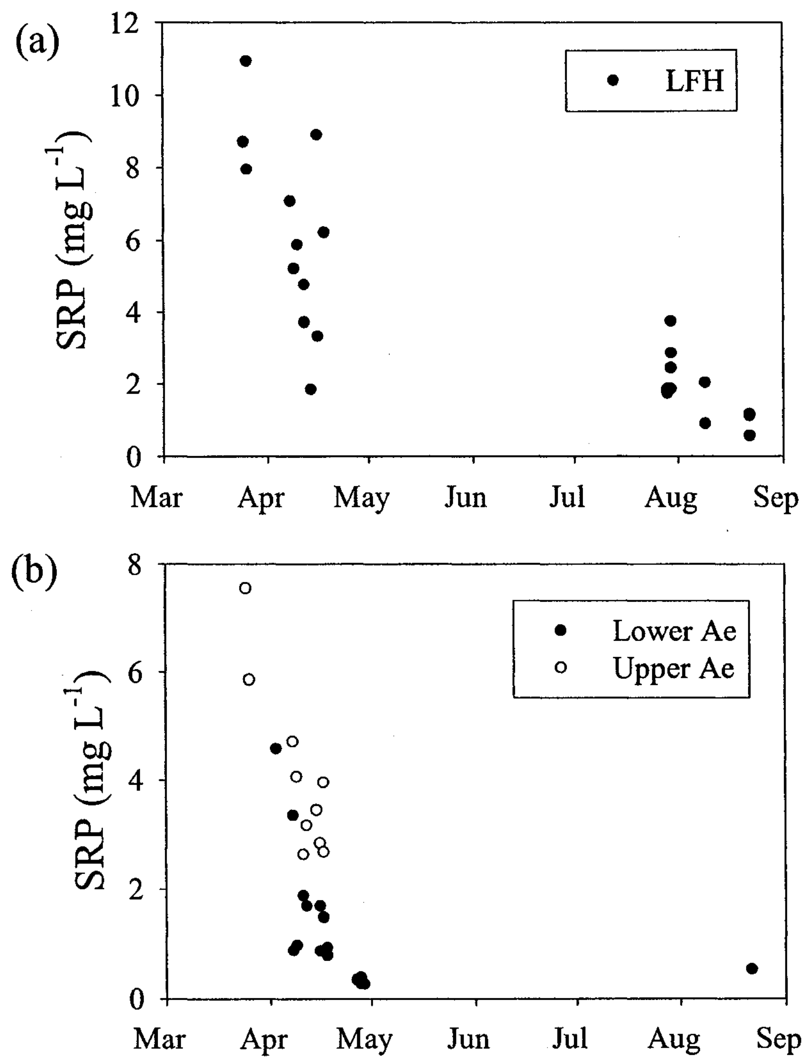


Figure 28: *In situ* soluble reactive phosphorus vs time for a) LFH, b) upper and lower Ae

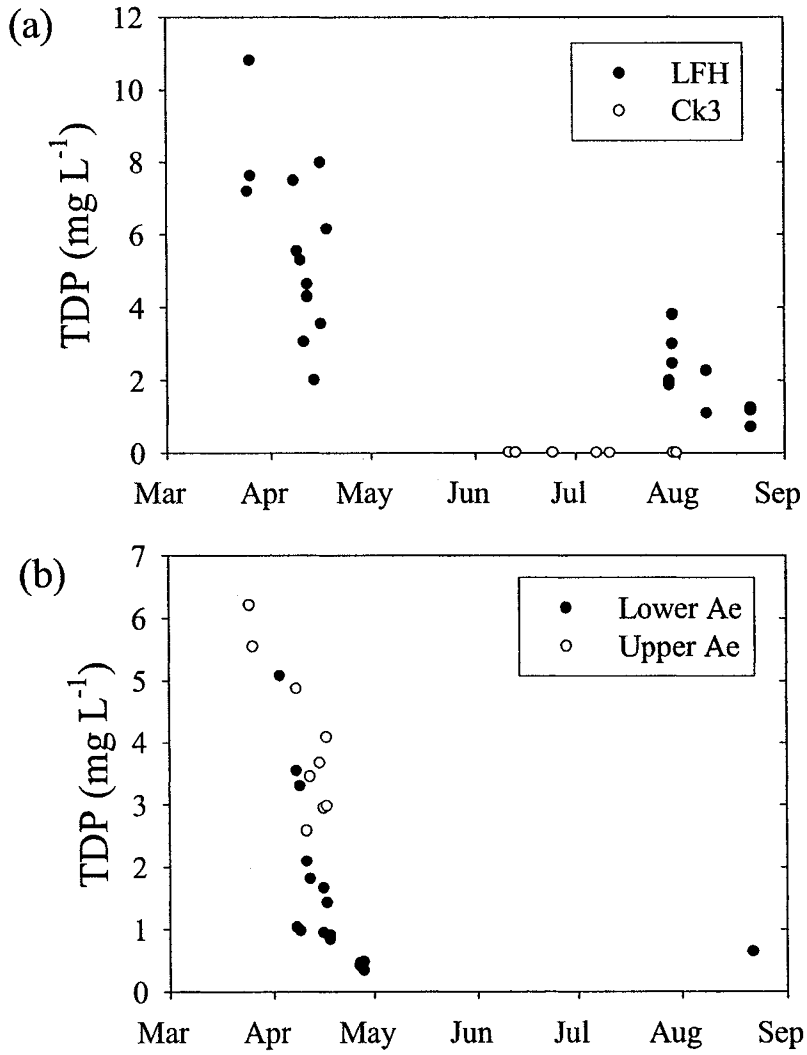


Figure 29: *In situ* total dissolved phosphorus vs time for a) LFH and Ck3, b) upper and lower Ae

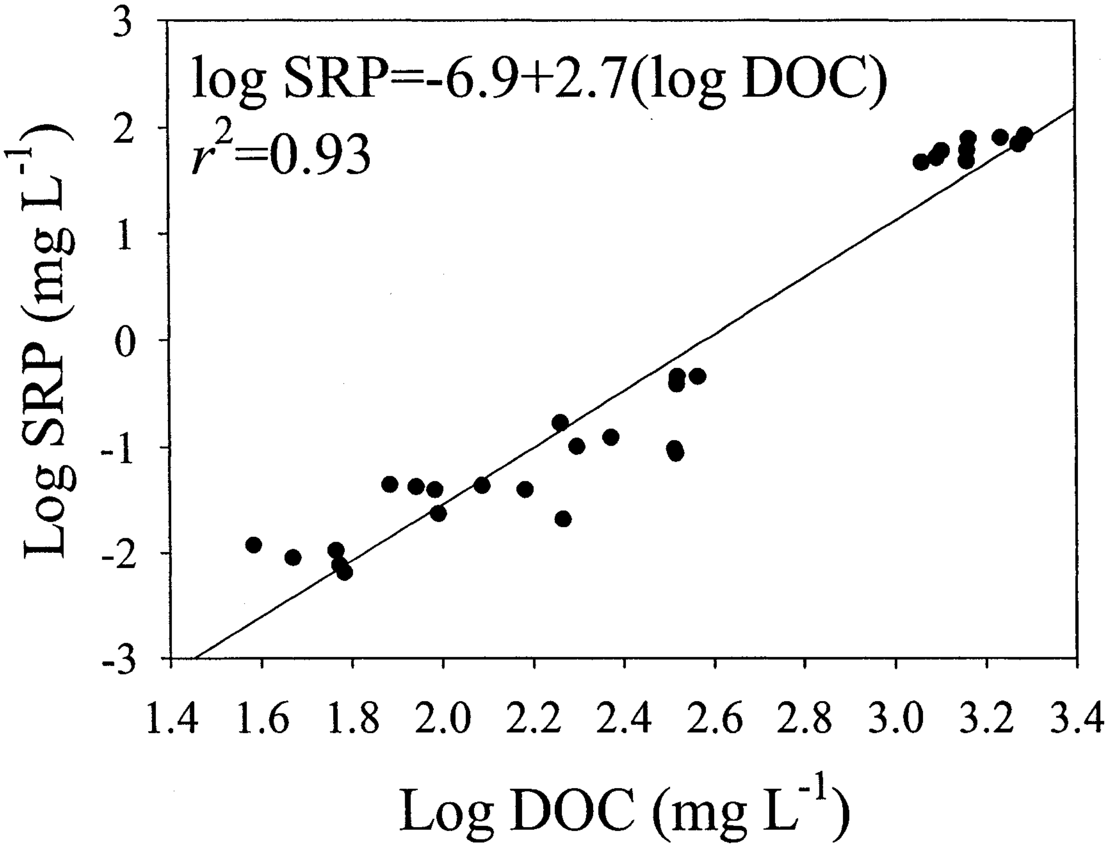


Figure 30: Relationship between dissolved organic carbon (DOC) and soluble reactive P (SRP)

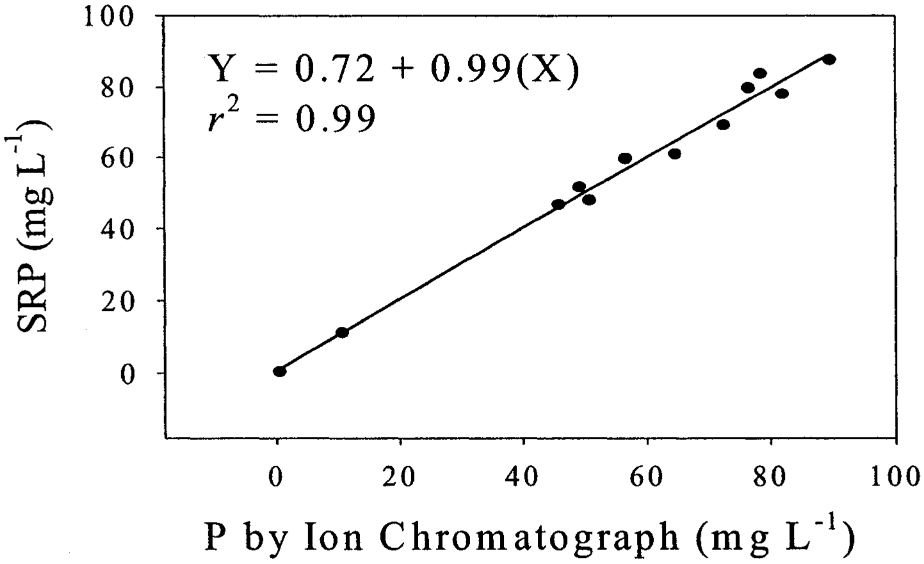


Figure 31: Relationship between ion chromatograph P and soluble reactive P (SRP)

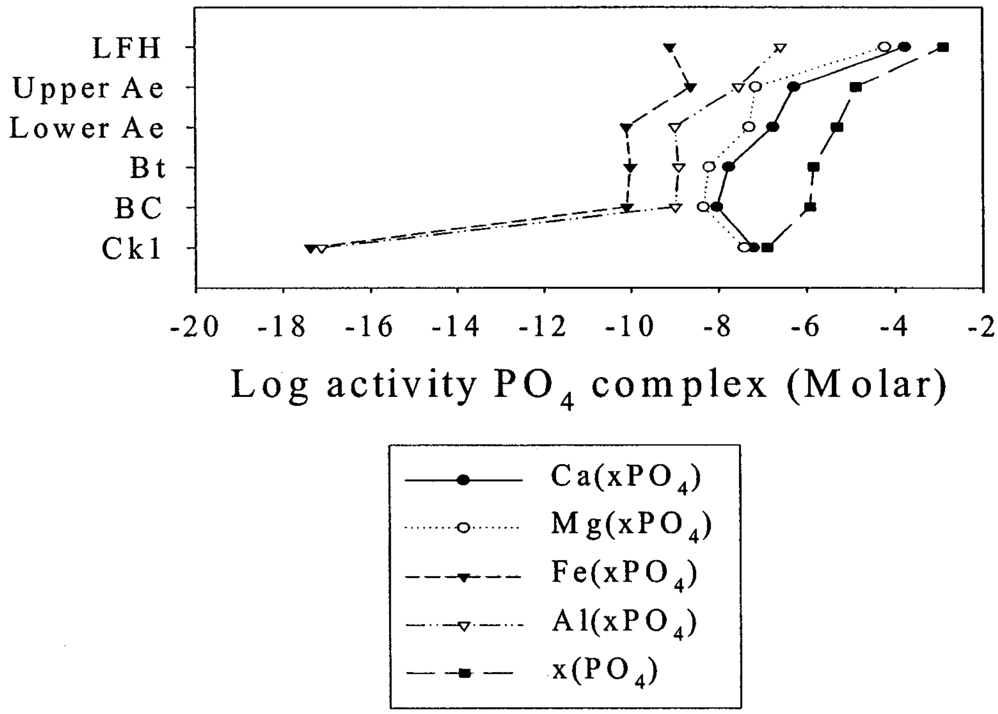


Figure 32: Distribution of phosphate complex activity by horizon as estimated by GEOCHEM ($x = H_n$)

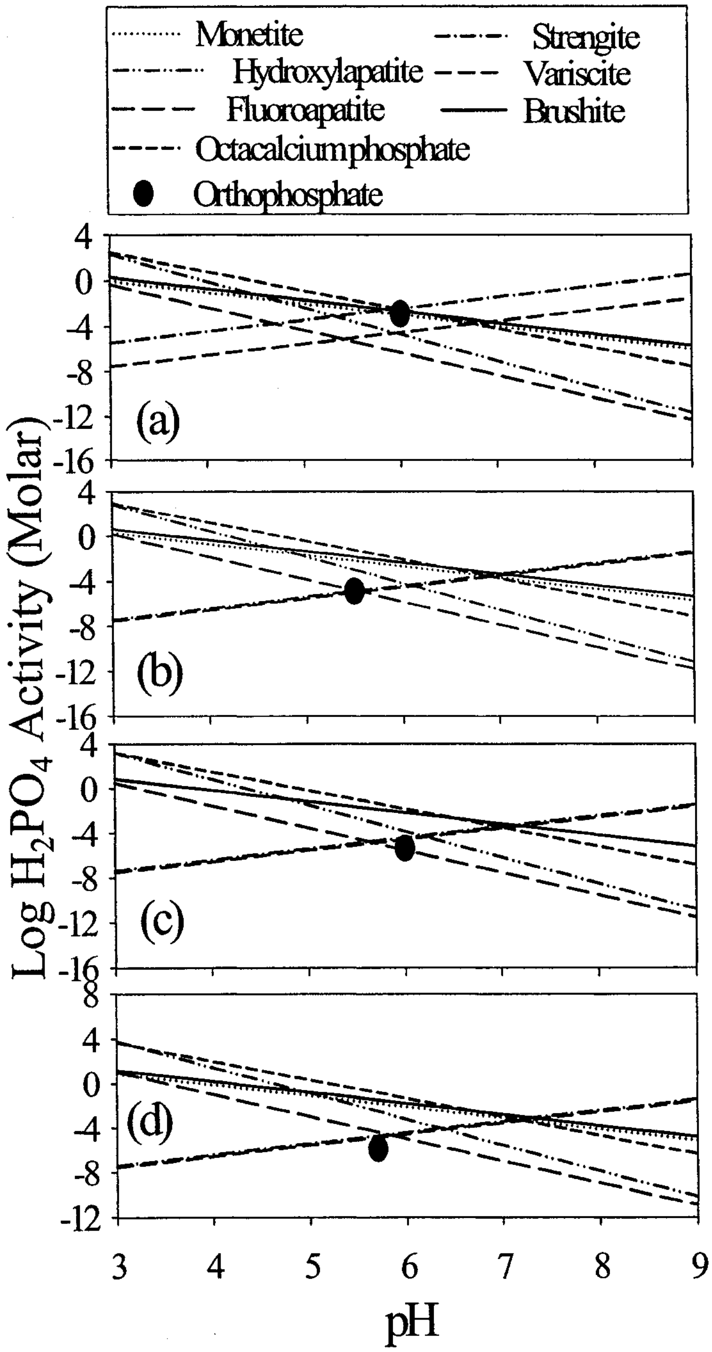


Figure 33: Stability diagram for phosphate as a function of horizon and pH: a) LFH, b) Upper Ae, c) lower Ae, d) Bt

## Chapter 6

## Effect of forest harvesting on potential phosphorus movement

### *Introduction*

The potential for forest harvesting to increase phosphorus export to streams and lakes in the Boreal Plain Ecozone has only recently received attention (Prepas *et al.* 2001a) despite strong earlier evidence of phosphorus limitation to aquatic ecosystem productivity (Schindler and Fee 1974; Prepas and Trew 1983). Many lakes in the region are eutrophic due to internal phosphorus loading (Prepas and Vickery 1984), so forest disturbances that increase phosphorus inputs from watersheds will exacerbate concerns with water quality. Despite great complexity in soil stratigraphy, flat topography, and highly variable precipitation that make prediction of forest disturbance on water quality difficult, harvest has been found to increase total phosphorus concentrations over the short term in regional lakes (Prepas *et al.* 2001a). However, the source of P reaching surface waters is poorly understood. It is important therefore to understand how different soil types and landforms contribute P to surface water in the watershed.

Forest harvest commonly increases runoff to streams and lakes. Clear cut harvesting in paired catchments in the NE USA increased annual runoff for a few years after harvest when a sufficient proportion of the watershed was affected by the disturbance (e.g., at least 25% of basal tree area removed) (Hornbeck *et al.* 1993). Much of the increased flow occurred shortly after harvest under expanded baseflow conditions, and was accompanied by wetter soils. Regrowth of vegetation gradually reduced water yield to background levels within 10 years (Hornbeck *et al.* 1993). Clayton and Kennedy (1985), working in Idaho, USA, found no increase in phosphorus transport in streams draining steeply sloping terrain when 23% of the watershed was disturbed. In coarse textured soils in northern Saskatchewan, clear cutting increased snowmelt and summer runoff by 5 and 2 times, respectively the year after harvest (Kachanoski and De Jong 1982). Snowmelt runoff at both Hubbard Brook and Marcell Experimental forests in the NE USA remained equal after harvest but peak flow occurred earlier (Hornbeck *et al.* 1993). Increased runoff implies that the phosphorus export will increase in proportion to runoff unless factors that control phosphorus concentration in solution are also affected by harvesting.

Many watershed scale studies in regions outside the Boreal Plain have found that harvesting seldom increases phosphorus export. At Hubbard Brook, clear cutting increased the export of nitrogen and some cations and anions (Bormann *et al.* 1968). While phosphorus release from fine decomposing roots was important to cycling (Fahey *et al.* 1988), the lack of increased phosphorus export was attributed to very strong sorption reactions that retained phosphorus in the mineral soil (Yanai 1991). Acidic soils dominated by iron and aluminum oxyhydroxide mineralogy can maintain very low solution P concentrations because of these sorption reactions (Ryden and Pratt 1980). In Wales, harvesting did not change the flux of soluble phosphorus entering streams from watersheds with acidic iron and aluminum oxide dominated soils (Stevens *et al.* 1995). In western Oregon, on soils of volcanic origin, soluble phosphorus did not increase after clear cut harvest (Martin and Harr 1989), probably because of the pattern of subsurface water movement through soil with very high phosphorus sorption capacity (Zeng *et al.* 2002). At sites with deciduous forest in Newfoundland, because of acidic soil conditions (Humo Ferric Podsols and Orthic Gleysols), harvesting in general did not detectably change the leaching of phosphorus (Titus *et al.* 1998). These examples illustrate that soils will be effective at retaining phosphorus only if their mineralogy favors relatively fast sorption reactions and if subsurface flow occurs. In Amazonia, forest removal and subsequent burning increased yield of soluble reactive phosphorus by a factor of 4 and total phosphorus (particulate and dissolved) by a factor of 8 from watersheds with acidic and highly weathered soils. Effects were attributed to increased runoff in surface pathways and to more erodible soil conditions (Williams and Melack 1997). The low phosphorus sorption capacity of the coarse textured soils meant that soil solution phosphorus concentration was probably controlled by the very slow rate of primary mineral weathering. The general dominance of subsurface flow in temperate forest environments (Anderson and Burt 1990) coupled with the often more weathered soil conditions in forests (Anderson 1988) probably explains why dissolved phosphorus export is often unaffected by harvest. However, because soil formation began only after that latest glacial period, the Boreal Plain ecozone has relatively unweathered soils (Pawluk 1961; Pawluk and Lindsay 1964; St. Arnaud and Whiteside 1964; Howitt and Pawluk 1985) with a very low iron and aluminum mineral component. Phosphorus is generally depleted from upper horizons of Gray Luvisols despite limited weathering (Xiao *et al.* 1991).

Complicating the issue of what harvesting does to phosphorus flux are the uncertainties of where, when, and how streamflow is generated in the watershed. The processes that generate runoff such as overland flow or subsurface flow and the parts of the watershed that contribute runoff will also affect the amount and kinds of phosphorus reaching the stream. Particulate phosphorus dominated export under storm flow conditions in both conifer-dominated (Munn and Prepas 1986) and aspen-dominated catchments (Cooke and Prepas 1998) on the Boreal Plain. Additionally, dissolved phosphorus concentrations in the soil solution were depth dependent, with very high concentrations found where shallow groundwater intersected soil horizons with high organic matter content (Evans *et al.* 2000) and concentrations similar to that found in baseflow associated with deeper groundwater (Shaw *et al.* 1990). Soluble reactive phosphorus concentrations decreased 3 to 4 orders of magnitude between the LFH horizon and the groundwater zone on a recently harvested Gray Luvisolic hillslope (Whitson *et al.* 2003c). All common Boreal Plain soils including dominant Luvisolic and Organic soils (Ecological Monitoring and Assessment Network 1996) accumulate organic matter at the surface, albeit it to different thicknesses. Peatlands, which are highly correlated with phosphorus export rates (Prepas *et al.* 2001b) have very shallow water tables whereas Gray Luvisols have much deeper water tables and are vertically stratified in terms of texture and structure such that profiles show pronounced anisotropic hydraulic conductivity (Coen and Wang 1989; Whitson *et al.* 2003a). Under suitable precipitation and antecedent moisture conditions, Gray Luvisols are expected to promote lateral subsurface flow (Whipkey and Kirkby 1978; Whitson *et al.* 2003b). Snowmelt runoff will favor near-surface flowpaths because of reduced hydraulic conductivity under frozen soil conditions (Granger *et al.* 1984) and will ensure that phosphorus in upper horizons is vulnerable to loss. Snowmelt is of particular importance in Boreal environments for runoff generation, and harvesting can influence patterns of snow accumulation (Berry and Rothwell 1992), and soil heat flux (Standish *et al.* 1988). Forest disturbance reduced infiltration rates in Gray Luvisols due to soil compaction (Whitson *et al.* 2003a). Surface crusting in the poorly aggregated upper mineral soil horizons will lead to a greater amount of flow through overlying organic horizons. Harvesting would also be expected to promote organic phosphorus mineralization because of higher soil temperatures. My general model of phosphorus transport of Gray Luvisols on the Boreal Plain views shallow flow through the LFH horizon during frozen soil conditions, and flow through Ae horizons during non-frozen conditions.

***Objectives and Hypotheses***

Chapters 2 through 5 of this thesis have explored the process of P movement in Gray Luvisolic soil profiles. Out of this work comes a clearer picture of how Gray Luvisols have lost so much P over their genesis (Xiao *et al.* 1991), and how P movement is likely to respond to forest harvest. There is potential for shallow flow through the LFH horizon during frozen soil conditions providing underlying air-filled porosity is minimal, a situation requiring very moist soil conditions. Interflow along the interface of the Ae and Bt horizons while expected in undisturbed conditions, seems to diminish as the result of compaction of the Ae horizon. While there is anisotropy within the Ae horizon such that lateral flow is encouraged over vertical, there were no differences in vertical hydraulic conductivity between the silt loam textured Ae horizon and the clay textured Bt horizon at this site. Very high dissolved, inorganic P concentrations in LFH and to some extent the upper Ae horizon potentially could translate into significant downslope movement of P along these shallow interflow pathways as compared to slower transport along the groundwater zone path. The findings outlined in Chapters 2 to 5 were used to develop a larger scale study of hillslopes with a greater range of topographic conditions to explore inter-hillslope differences in the processes affecting P movement on this important soil type. The objective of these studies was to infer dissolved P export in typical upland forest soils occurring on a range of topographic conditions typical of the region from measurements of key related soil variables. I viewed the most important variables as frozen soil conditions, shallow interflow, antecedent moisture conditions, and phosphorus availability. These variables were used as indicators of how these hillslopes would respond to forest harvest in terms of phosphorus transport. To this end, I used plot-scale studies on hillslopes with a range of topographic characteristics, including different slope gradients, aspects, and shapes (Figure 34).

Harvesting was hypothesized to:

- increase the availability and concentration of phosphorus in the soil horizons most likely to generate runoff in Gray Luvisols.
- result in wetter soil conditions
- accelerate soil thaw in the spring
- increase the amount of subsurface lateral flow towards lower topographic

conditions.

*Materials and Methods*

Field work was carried out approximately 250 km north of Edmonton, Alberta, Canada, within the Boreal Mixedwood Ecoregion (Strong and Leggat 1992). Glacial deposits of the study region were underlain by upper Cretaceous bedrock consisting primarily of alternating beds of non-marine shale and sandstone (Green 1972). Topography was typical of morainal landforms with slope gradients as high as 30% and slope contours varying from convex to concave. Soils in upland locations were primarily Gray Luvisols while poorly drained areas were occupied by soils of the Organic Order. Upland forest species consisted of trembling aspen (*Populus tremuloides* Michx.), white spruce (*Picea glauca* Voss), jack pine (*Pinus banksiana* Lamb.) and balsam poplar (*P. balsamifera* L.) while peatlands supported stands of black spruce (*P. mariana* [Mill.] BSP.). Mean July and January temperatures for the 1950 to 1980 period at nearby Smith, Alberta (40 km distant) were 15.6 and -19.6 °C respectively (Environment Canada 2002a) while mean annual precipitation was 502 mm, of which 29% was snow. Harvest took place in January 1997, when snow was approximately 40 cm deep and while soils were frozen. Harvest methods involved both cut-to-length and whole-tree-length methods using a combination of forwarders, grapple skidders, and feller bunchers. Most conifer-dominated areas received no post-harvest silvicultural treatment other than reforestation with white spruce seedlings. Areas previously dominated by aspen were allowed to regenerate naturally.

Study plots were selected within an area of 2 km radius close to existing trails. The great variability in topographic characteristics in this morainal landscape made replication difficult and forced us to select sites with a range of slope gradient, shape, aspect and vegetation characteristics. As a solution, treatment sites were matched with reference sites that were similar in terms of topography and vegetation. Sites, each approximately 100 $m^2$ in area, were chosen at mid to upper slope position on convex or concave shaped hillslopes. Eleven pairs of reference-treatment sites were selected, and were numbered sequentially from 46 to 69. Slope gradient was measured with a clinometer. Soil profiles were classified to the subgroup level of the Canadian System of Soil Classification (Agriculture Canada Expert Committee on Soil Survey 1987) and to the series level (Knapik and Brierley 1993) with reference to parent material and landform characteristics.

Soil samples were collected by horizon from one profile at each site, air dried, and ground to pass a 2-mm size sieve and stored at 20 °C until analysis. Samples were analyzed for particle size distribution by the hydrometer method (Sheldrick and Wang 1993). pH was measured in 0.01 M $CaCl_2$ solution (1:2 soil to water by weight) with a Mettler 925 pH meter and electrode. Saturated paste extracts (Janzen 1993) were obtained for phosphorus analysis from Ae and LFH horizons from a subset of the 22 sites. After suctioning samples through 0.45-µm filters, soluble reactive phosphorus concentration was determined by the ascorbic acid method (Murphy and Riley 1962) with absorbance read at 885 nm by spectrophotometer.

Soil samples were also collected from the LFH and Ae horizons at 2 or 3 randomly located points at the plot four times over the study period for nutrient analysis. Samples were air dried and ground to less than 2 mm, and stored at 20 °C until analysis. Extractable phosphorus was determined with a solution of 0.03 N ammonium fluoride/sulfuric acid, and a 1:5 soil to solution suspension subjected to 10 minutes of shaking, with phosphorus concentration determined by Technicon autoanalyzer.

A tracer, calcium bromide, was used to estimate differences in the extent of subsurface water flow between treatments. Prior to application, 2.0 m by 2.0 m plots were cleared of vegetation to ensure that all tracer reached the land surface. Bromide was dissolved in approximately 12 L of water and applied with a hand bucket sprinkler a rate of 20.0 g $m^{-2}$ in August 1998, during an extended period of dry weather. Plots were sampled in June and October of the following year. Vegetation did not reestablish on the plots during the remainder of the study. Soil was obtained by hand auger in increments of the LFH horizon, 0 to10, 10 to 20, and each 20-cm increment thereafter to a 160-cm depth at four locations in the plot. Samples within a depth increment were combined, then air dried, ground to less than 2 mm, and stored at 20 °C until extraction. Bromide was extracted with 0.005 M $CaCl_2$ solution after 1 hour of vigorous shaking. Extracts were centrifuged, suctioned through GF/F filters, and stored at 5 °C. Bromide concentration was determined with a Dionex Ion Chromatograph equipped with an AS-9 column, allowing a detection limit of 0.1 mg $L^{-1}$. Sample concentrations were converted from mass to volumetric basis with soil bulk density measurements taken in increments to 160 cm at several of the sites (Whitson *et al.* 2003b). The quantity of bromide at a plot on an areal basis was calculated from the summation of the quantity within each depth increment.

Moisture content was determined by sampling soil in increments to a depth of up to 160 cm with a hand auger on 4 occasions. The LFH was sampled separately, whereas underlying mineral soil was sampled from 0 to 10, 10 to 20, and each 20 cm thereafter. The number of sample points per site varied by sample event. During the first sampling, a single profile was taken as part of temperature probe installation, while during the third, two profiles 10 m apart were obtained. During the second and fourth events, four profiles from the 4 $m^2$ square bromide plots were sampled and combined within each increment. Soil was weighed before and after drying at 105 °C to determine mass water content, which was converted to volumetric water content with bulk density values.

Soil temperature was measured to estimate the date of soil thaw. Type T thermocouple wire of equal lead length was installed at 7 depths, +5, -10, -25, -45, -65, -85, and -105 cm relative to the mineral soil surface. Temperature was measured during daylight hours several times weekly from late March to mid-June with a manual thermocouple thermometer. Thermocouples had an accuracy of ± 0.5 °C. Measurements were collected from individual treatment and reference pairs within a 1 h time period to minimize temperature change. Soil thaw was estimated from visual inspection of the temperature *versus* date data. A given depth was assumed to be frozen if the temperature remained near 0 °C for an extended period, while thaw was assumed once successive temperature measurements stayed above 0 °C. Air temperature was also recorded at one of the harvested sites with a Campbell Scientific CR10X datalogger.

Choice of technique for statistical evaluation depended on the assumptions of sample independence. Soil temperature and moisture content were expected to be influenced by topographic characteristics, such as through differences in energy flux with aspect. Likewise, tracer movement would reflect differences in hydraulic gradient due to slope. Treatment effects for these topographic dependent variables were determined by paired *t*-test. Soil chemical parameters such pH and extractable phosphorus were not related to topographic characteristics and so were analyzed with the assumption of sample independence. Treatment effects on pH, soluble reactive and extractable phosphorus were determined by an independent samples *t*-test with significance less than or equal to 0.05.

***Results***

While forest stands at sites ranged from aspen to white spruce dominated, most were best described as mixedwood. Slope gradients at plots ranged from 6 to 25% (Table 20), with an average of 17% for both treatment and reference sites. Ae horizons were usually silt loam, reflecting eluviation of clay, whereas B horizons were typically silty clay-loam to clay textured. Parent materials consisted of till that was usually clay to clay loam in texture. Major soil horizons were similar between the two treatments in terms of clay ($0.08<P<0.16$), silt ($0.46<P<0.80$), and sand content ($0.06<P<0.91$) (Table 21). Horizon thicknesses were also very similar among the treatments. Mean LFH horizon thickness was 9 cm under forest and 8 cm in harvested areas. Average Ae horizon thickness was 16 and 14 cm at forested and harvested sites, respectively. Soils were classified as Orthic Gray Luvisols at 18 sites, as Dark Gray Luvisols at 3 sites, and as Gleyed Gray Luvisol at one site. Site 47, an Orthic Gray Luvisol, was eroded, with a blocky-structured AB horizon (probably a weathered Bt horizon) immediately beneath the forest floor. Two of the sites classified as Dark Gray Luvisol (Sites 68 and 69) were situated just north of a lake on a south-facing slope.

Precipitation for the period of study was only about 50% of the long term average (Whitson *et al.* 2003b). May 1 to October 31 precipitation was 168 mm in 1998 and 179 mm in 1999, while a nearby harvested slope with an eastern aspect had a snow water equivalent of 87 mm in February, 1999. Very dry weather conditions in 1999 meant that runoff at nearby Driftwood River was reduced to 9% of the long-term mean annual yield (Environment Canada 2002b). Despite the dryness, aspen re-colonized the harvested sites during the three post-harvest years (1997 to 1999), typically reaching 2 m or more in height and attaining high density even in areas where white spruce had been the dominant tree prior to harvesting.

Soil nutrient conditions in the litter layer differed between treatments. In the LFH horizon, pH was higher at harvested (5.8±0.2) than forested (5.2±0.2) sites ($P$=0.05). There was no detectable difference in pH between forested and harvested sites for the Ae (4.7±0.1), Bt

(4.7±0.1), and BC (4.4±0.1) horizons ($0.46<P<0.81$). Water soluble phosphorus in the LFH horizon was significantly lower at harvested sites (41±9 mg $L^{-1}$) than at forested sites (74±9 mg $L^{-1}$) ($P=0.03$) but was not detectably different in the Ae horizon (0.13±0.03 mg $L^{-1}$) ($P=0.93$). On two of four occasions extractable phosphorus was significantly lower ($P<0.02$) at harvested sites than at forested sites in the LFH horizon (Table 22). However, differences in extractable phosphorus between forested and harvested sites were not detectable in the Ae horizon ($0.22<P<0.95$). Extractable phosphorus also varied temporally within reference and treatment sites (Table 22).

There were detectable treatment effects on soil hydraulic properties, particularly in terms of subsurface flow as revealed by the bromide data. Bromide recovery was higher ($P=0.03$) under forested sites by the second sampling. An average of 2207 ± 984 mg $m^{-2}$ more bromide was found at forested than at harvested sites. The bromide plume remained above the 80-cm depth (Figure 35), with the plume slightly deeper under the forested sites than in the cutblock. However, there was no relationship between slope gradient and bromide recovery at either forested or harvested sites. Soil moisture was lower at forested sites in the 0 to 80 cm zone ($P<0.04$) in July 1999, and for the 0 to 20 cm ($P<0.02$) increment in October 1999 (Figure 36).

Temperature of upper soil increments followed air temperature trends better than lower soil depths (Figure 37a-h). There was a consistent trend towards higher soil temperature at harvested sites (Figure 37b-h). Over the period of measurement, soil temperature at harvested sites was greater than that at forested sites on 8 occasions in the LFH, 3 occasions at 10 cm depth, once each at 45, 65, and 105 cm depth, and twice at 85 cm depth (Table 23). In spite of trends towards warmer conditions, there were no detectable differences in the date of thaw for any soil increment between forested and harvested sites. The LFH horizon thawed approximately 2 weeks prior to the upper mineral soil while subsoils thawed typically in 1 to 2 weeks following thaw in the upper mineral soil increment for both treatment and reference sites (Figure 38). The influence of aspect on profile thaw date was however readily apparent (Figure 39). Profiles of northerly aspect typically melted in the latter half of May, whereas those with southerly aspect began thawing in mid April.

***Discussion***

Field evidence did not support my hypotheses of greater extractable or water soluble phosphorus after harvest. Similar results were reported by Schmidt et al. (1996), who found that forest harvesting and site preparation greatly reduced the phosphorus availability in the LFH horizon at two sites also with Gray Luvisolic soils in western Alberta. However, an opposite trend was found in the southern Boreal forest of Quebec, where phosphorus availability declined with time since fire, reaching levels in 231-year old stands of about half those in 47-year old stands (Pare *et al.* 1993). Extractable phosphorus measurements can be viewed as an index of potential solution concentration (Pote *et al.* 1996). Reduction of extractable phosphorus content in the LFH horizon after harvest at my sites might be due to increased microbial immobilization, or to a switch towards younger vegetation with a larger phosphorus demand. Differences may also be explained by the reduced calcium-phosphate solubility with pH increase (Lindsay and Moreno 1960; Lindsay 1979; Lindsay 1981). Reduced available P could also reflect the importance of rates of throughfall or litterfall, both of which may differ after harvest and subsequent regrowth of fast growing aspen. Reductions in these phosphorus indices suggest that dissolved P export from Gray Luvisols will not increase in proportion to runoff changes after harvest.

Use of the tracer study as an index of lateral convective flow was confounded by possible vegetation uptake of the bromide. Although vegetation was removed from plots prior to tracer application, roots in the Ae horizon and lower LFH remained and may have absorbed bromide along with water. If we assume that vegetation uptake was either minimal or at least equal in the forested and harvested sites, then the greater recovery of bromide from forested sites can be interpreted as different rates of convective flow between the two treatments. Lower recovery of bromide at the harvested sites implies greater lateral, convective flow. Given that soil water contents were below saturation most of the time, convective flow, if it occurred at all, was happening in unsaturated conditions. The greater depth of the bromide plume in forested sites may reflect greater evapotranspiration or different precipitation rates between the two treatments. Although the previous winter snowpack was not measured, lower snow water equivalent in the harvested areas would have reduced spring infiltration and subsequent downward movement of the bromide plume. Reduced snow depth after harvesting may be a function of higher wind speeds and increased sublimation losses. The absence of a relationship between slope gradient and bromide loss is puzzling, given that the lateral Darcian moisture flux is controlled by the hydraulic gradient, but probably reflects spatial variability of hydraulic conductivity. At low soil moisture contents, high spatial variability of hydraulic conductivity (Warrick and Nielsen 1980) will make it difficult to detect any effect of slope gradient on lateral moisture flux.

Limited differences in moisture content provided some support for the hypothesis that harvesting produced wetter soils even in these arid conditions. Rapid aspen colonization and growth may quickly restore soil moisture levels to pre-disturbance values because of their high leaf area index. These results highlight the importance of replication to smooth out local variability in soil moisture content introduced by tree canopy effects on rainfall. Re-establishment of conifer plantations after harvest, through use of herbicide to reduce competition from deciduous species, could result in far wetter soils than I observed at these naturally regenerating aspen-dominated sites. Forest species composition changes have lead to unexpected changes in the water balance following harvest, due to inter-species differences in interception and transpiration properties (Hornbeck *et al.* 1993). The rapid regrowth of aspen at very high densities in mixedwood stands, an unintended consequence of harvest techniques and a natural component of Boreal forest succession, means that soil moisture conditions will quickly return to pre-disturbance values, with implications for the regional water balance and phosphorus transport.

Results of soil temperature measurements provide some support for the hypothesis of more rapid soil thaw after harvest than in forest. Trends towards higher soil temperature under harvested areas suggest that energy inputs are greater when the trees are removed.

LFH horizons in harvested areas were often 1 to 2 °C warmer than forested locations with equivalent topography. Although these temperature differences did not result in faster profile thaw on harvested sites, they would probably have positive consequences for forest productivity. Aspect had a marked effect on thaw rate. Topography will influence the intensity of snowmelt runoff, with southern aspects melting earlier (Carey and Woo 1998). Watersheds with a wide range of aspects would probably have lower peak flow during snowmelt than those with a restricted range of aspects, because the snowmelt and subsequent soil thaw would extend over a longer period of time at the former.

The overall effect of harvest on phosphorus export from these soils will depend on the balance between forces that increase export (runoff) with those that will tend to reduce it (decreased phosphorus solubility). High inter-annual variation in runoff in this region reflects similar variation in annual precipitation (Prepas *et al.* 2001a). The TROLS study (Prepas *et al.* 2001a), which evaluated changes to lakes following harvest in watersheds with more complex soil conditions, found that forest removal resulted in detectable increases in TP in thermally-stratified lakes during the first post-treatment year (high rainfall) but not in the second year (low rainfall conditions of 1998). Phosphorus export from Gray Luvisols is probably more dependent on the factors that control runoff than on those that control phosphorus solubility.

***Conclusions***

Harvesting decreased phosphorus availability in the forest floor but not the underlying mineral soil. There were trends towards warmer, wetter soils and increased lateral subsurface flow through upper soil horizons at harvested sites despite very dry weather. It appears that the net effect of harvest on phosphorus export from hillslopes with Gray Luvisolic soils will depend on the balance between increased runoff and decreased soil solution phosphorus concentration.

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Table 20: Topographic and soil textural properties of forested-harvested sites

| Pair No. | Slope Shape | Forested | | | | | Harvested | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | | Aspect | Slope Gradient % | Textural Class by Soil Horizon | | | Aspect | Slope Gradient % | Textural Class by Soil Horizon | | |
| | | | | Ae or Ahe | Bt or Btgj | BC | | | Ae or Ahe | Bt or Btgj | BC |
| 46-47 | convex | NE | 23 | SiL | C | C | N | 23 | CL | CL | CL |
| 48-49 | convex | SW | 16 | L | C | C | SW | 15 | SiL | C | C |
| 50-51 | convex | S | 15 | SiL | C | C | S | 13 | SiL | CL | CL |
| 52-53 | concave | E | 16 | SiL | C | CL | E | 14 | SiL | C | C |
| 54-55 | concave | SW | 6 | SiL | SiCL | CL | SE | 6 | SiL | C | C |
| 56-57 | convex | E | 14 | SiL | SiC | C | E | 13 | SiL | C | C |
| 58-59 | convex | N | 17 | SiL | SiC | SiCL | NW | 19 | SiL | SiCL | SiCL |
| 60-61 | convex | N | 21 | | C | C | N | 19 | SiL | CL | CL |
| 64-65 | convex | N | 25 | SiL | C | C | N | 25 | SiL | CL | CL |
| 66-67 | convex | N | 7 | SiL | C | C | NW | 8 | SiL | CL | L |
| 68-69 | convex | S | 19 | SiL | C | C | S | 20 | L | C | SiCL |

Table 21: Particle size distribution by horizon (mean ± 1*SE*) for all sites

| Horizon | Clay (%) | Silt (%) | Sand (%) |
|---|---|---|---|
| Ae | 15±1 | 67±2 | 18±2 |
| Bt | 42±1 | 35±2 | 23±2 |
| BC | 38±1 | 36±2 | 26±1 |

Table 22: Extractable phosphorus contents of soils (mg $kg^{-1}$) by horizon and treatment (mean ± 1*SE*)

| Sample Date | LFH Horizon | | Ae Horizon | |
|---|---|---|---|---|
| | Forested | Harvested | Forested | Harvested |
| August 1998 | 103 ± 10 | 79 ± 10 | 32 ± 6 | 29 ± 6 |
| October 1998 | 80 ± 7 | 59 ± 4* | 31 ± 4 | 27 ± 6 |
| May 1999 | 132 ± 6 | 96 ± 6** | 40 ± 6 | 40 ± 6 |
| June 1999 | 127 ± 8 | 112 ± 7 | 48 ± 7 | 38 ± 4 |

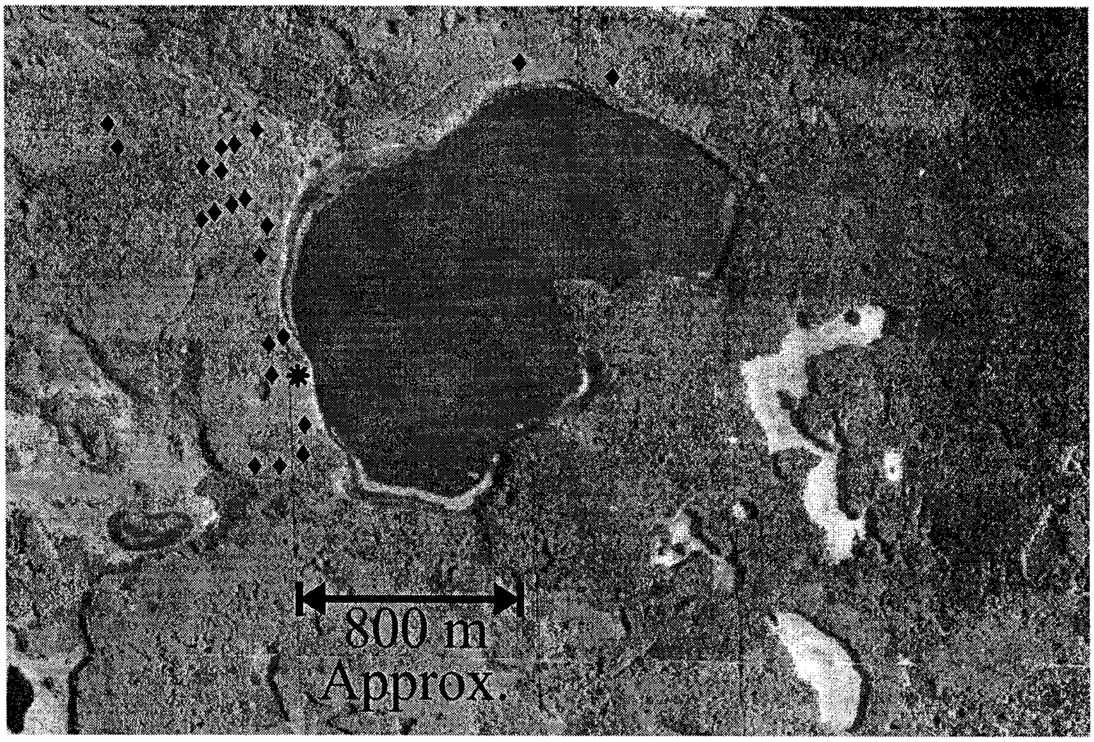
* $P = 0.02$

** $P = 0.01$

Table 23: Soil temperature difference†,†† between harvested and forested sites in spring 1999

| Date | | Temperature increment depth in cm relative to mineral soil surface | | | | | | |
|---|---|---|---|---|---|---|---|---|
| | | +5 | -10 | -25 | -45 | -65 | -85 | -105 |
| March 26 | °C | | | | +0.5 | +0.5 | | |
| | P | | | | 0.007 | 0.001 | | |
| April 3 | °C | | | | | | | +0.5 |
| | P | | | | | | | 0.005 |
| April 27 | °C | +1.1 | | | | | | |
| | P | 0.011 | | | | | | |
| May 6 | °C | +2.7 | +1.3 | | | | | |
| | P | 0.030 | 0.032 | | | | | |
| May 12 | °C | +1.2 | | | | | | |
| | P | 0.047 | | | | | | |
| May 13 | °C | +2.0 | +1.3 | | | | | |
| | P | 0.039 | 0.021 | | | | | |
| May 14 | °C | | +1.1 | | | | | |
| | P | | 0.019 | | | | | |
| May 16 | °C | +1.9 | | | | | | |
| | P | 0.019 | | | | | | |
| May 17 | °C | +1.4 | | | | | | |
| | P | 0.011 | | | | | | |
| May 26 | °C | +2.3 | | | | | | |
| | P | 0.007 | | | | | | |
| May 27 | °C | | | | | | | +0.6 |
| | P | | | | | | | 0.049 |
| May 29 | °C | | | | | | +0.9 | |
| | P | | | | | | 0.028 | |
| May 31 | °C | | | | | | +1.1 | |
| | P | | | | | | 0.049 | |
| June 9 | °C | +1.1 | | | | | | |
| | P | 0.041 | | | | | | |

† mean temperature difference in °C (harvested relative to forested sites) and significance of paired *t*-test
†† differences less than 0.5 °C were within the range of thermocouple error




✱ Replicated Hillslope Study

♦ Forest-Harvest Comparison Sites

Figure 34: Aerial view of study sites

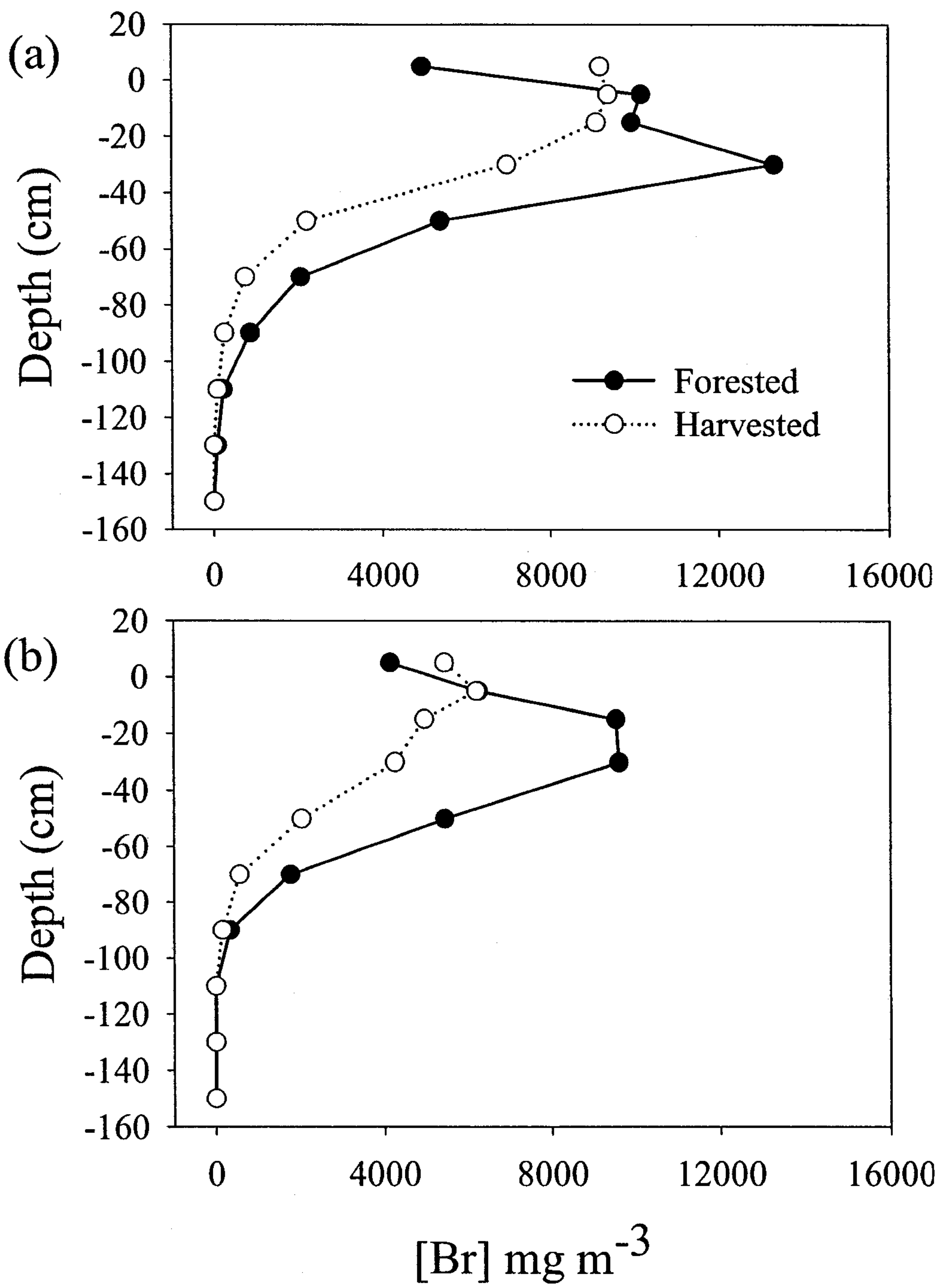



Figure 35: Mean bromide concentration *versus* depth for forested and harvested sites in a) spring and b) fall

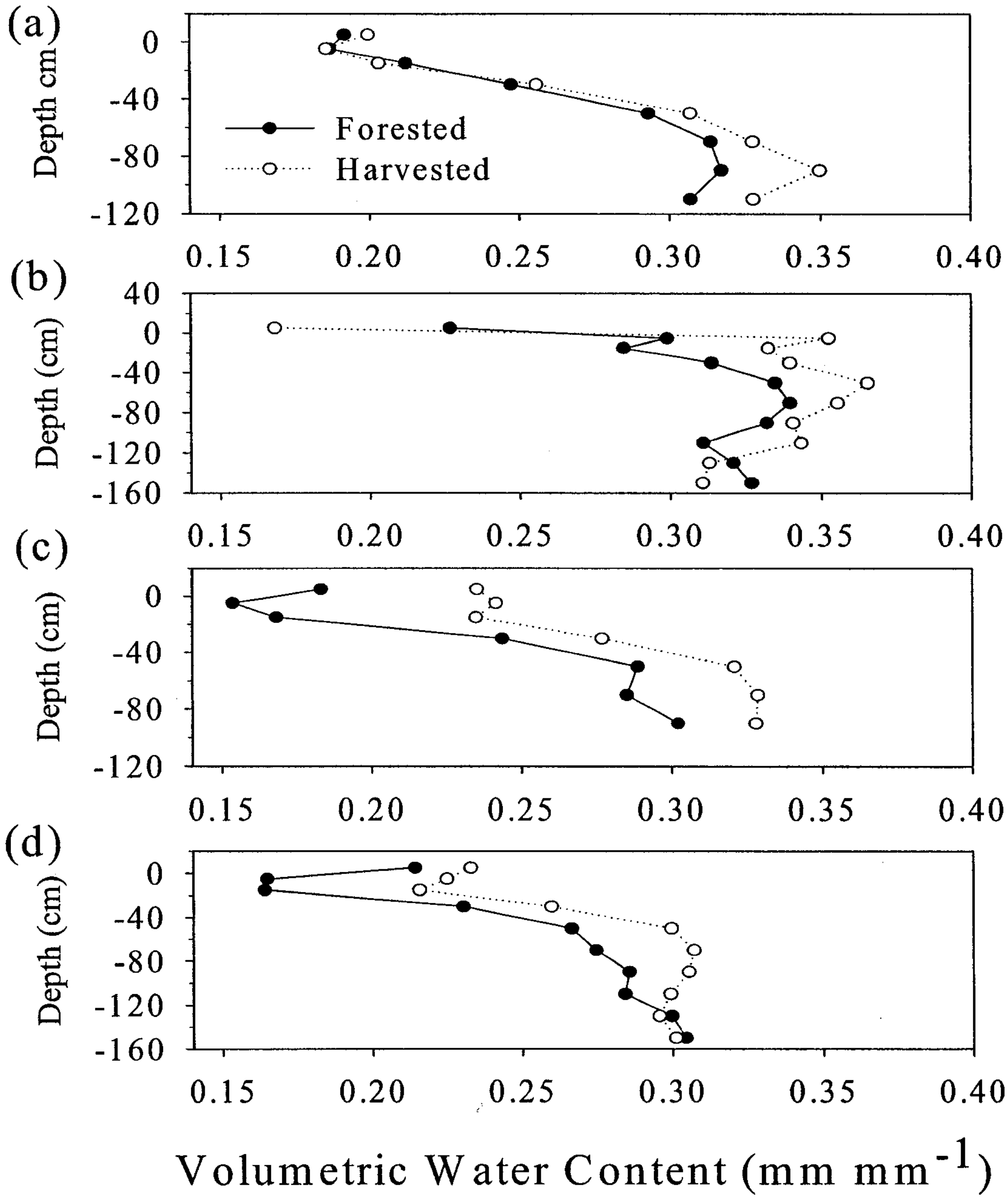
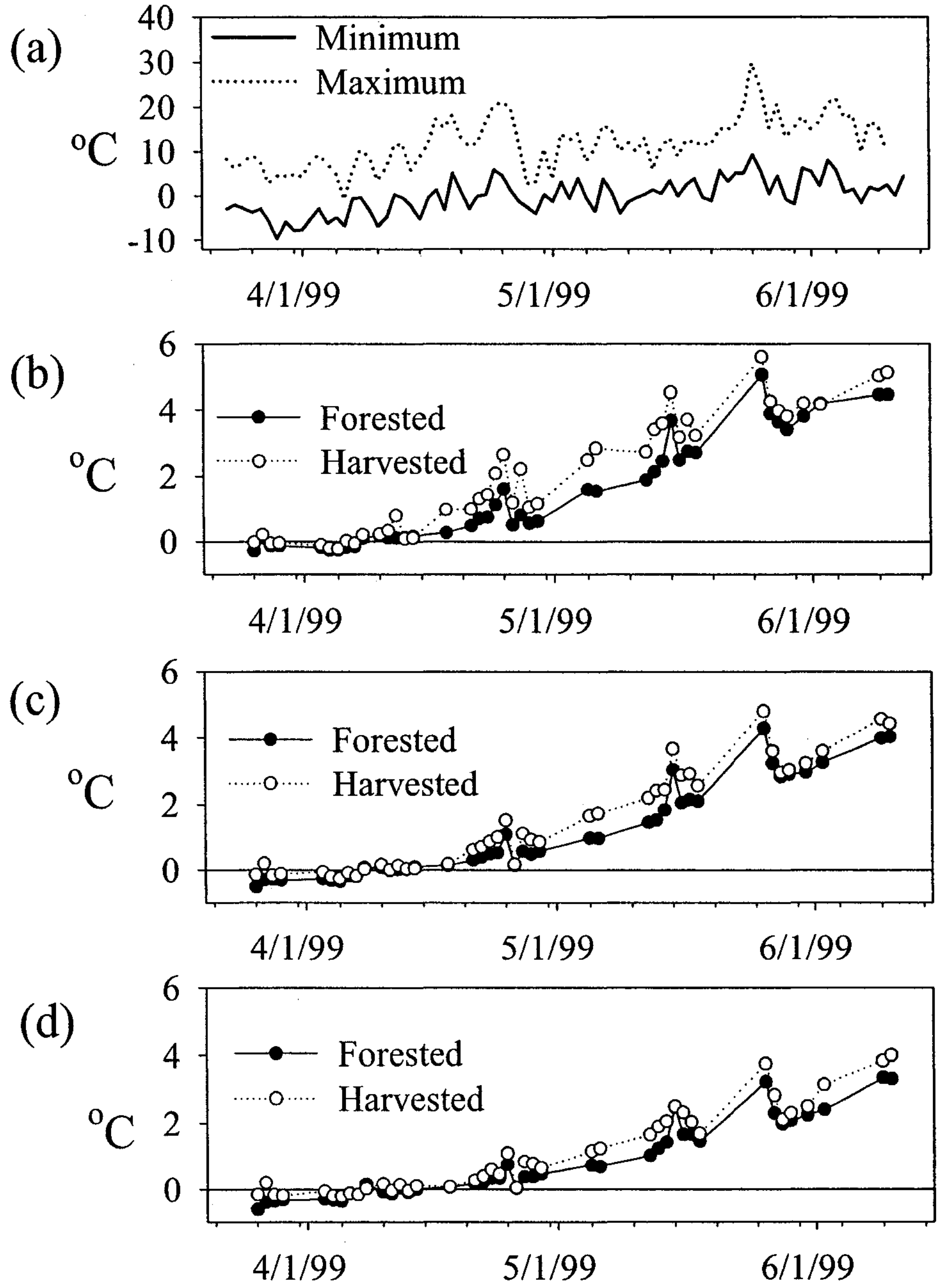



Figure 36: Mean volumetric water content (mm mm$^{-1}$) by depth from forested and harvested sites on 4 occasions a) August 1998, b) June 1999, c) July 1999, and d) October 1999



(a)
40
30
20
°C 10
0
-10
Minimum
Maximum
4/1/99
5/1/99
6/1/99
(b)
6
4
°C 2
0
Forested
Harvested
4/1/99
5/1/99
6/1/99
(c)
6
4
°C 2
0
Forested
Harvested
4/1/99
5/1/99
6/1/99
(d)
6
4
°C 2
0
Forested
Harvested
4/1/99
5/1/99
6/1/99

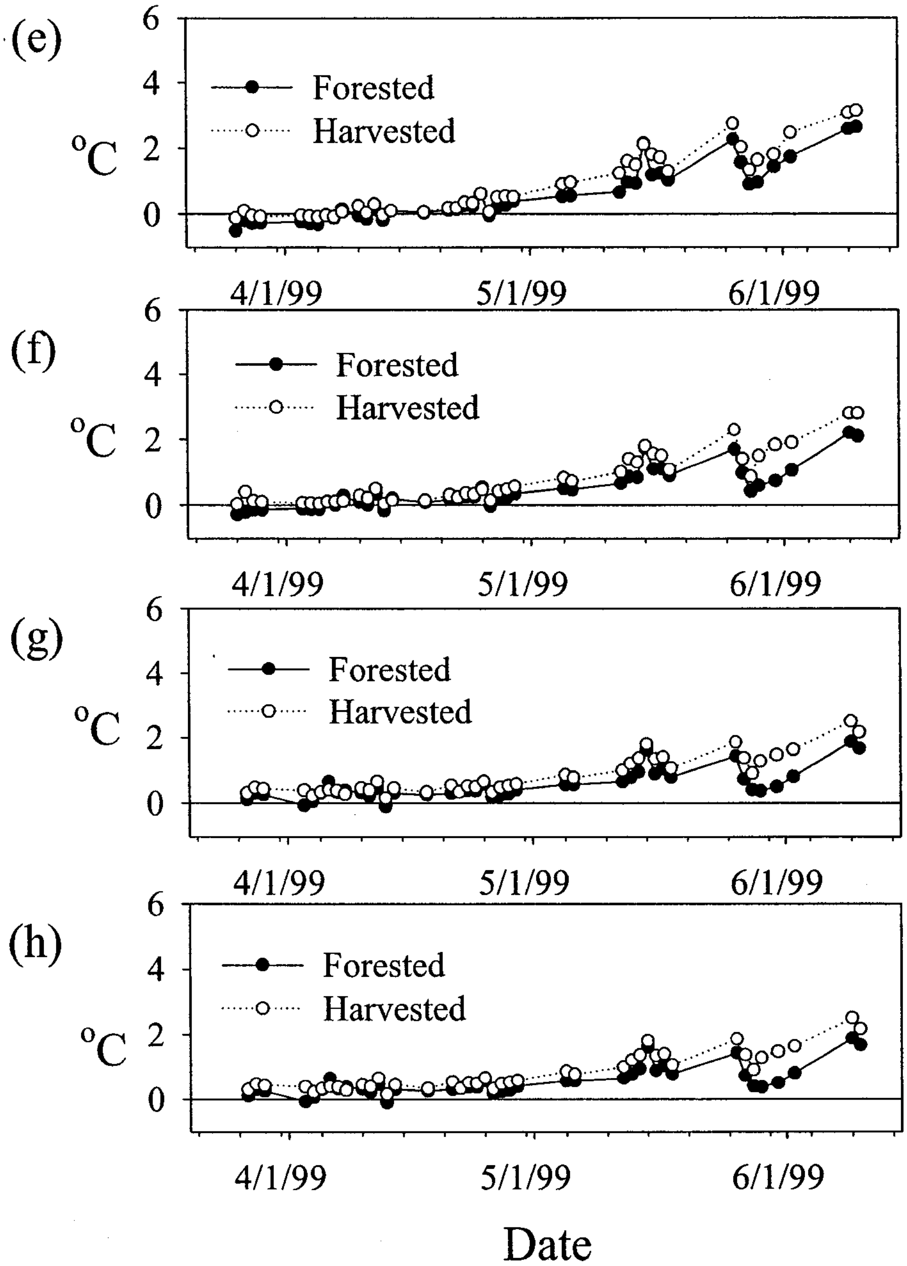



Figure 37: Daily air temperature (a) and mean soil temperature by depth over time at forested and harvested sites: b) LFH, c) 10 cm, d) 25 cm, e) 45 cm, f) 65 cm, g) 85 cm, and h) 105 cm depth

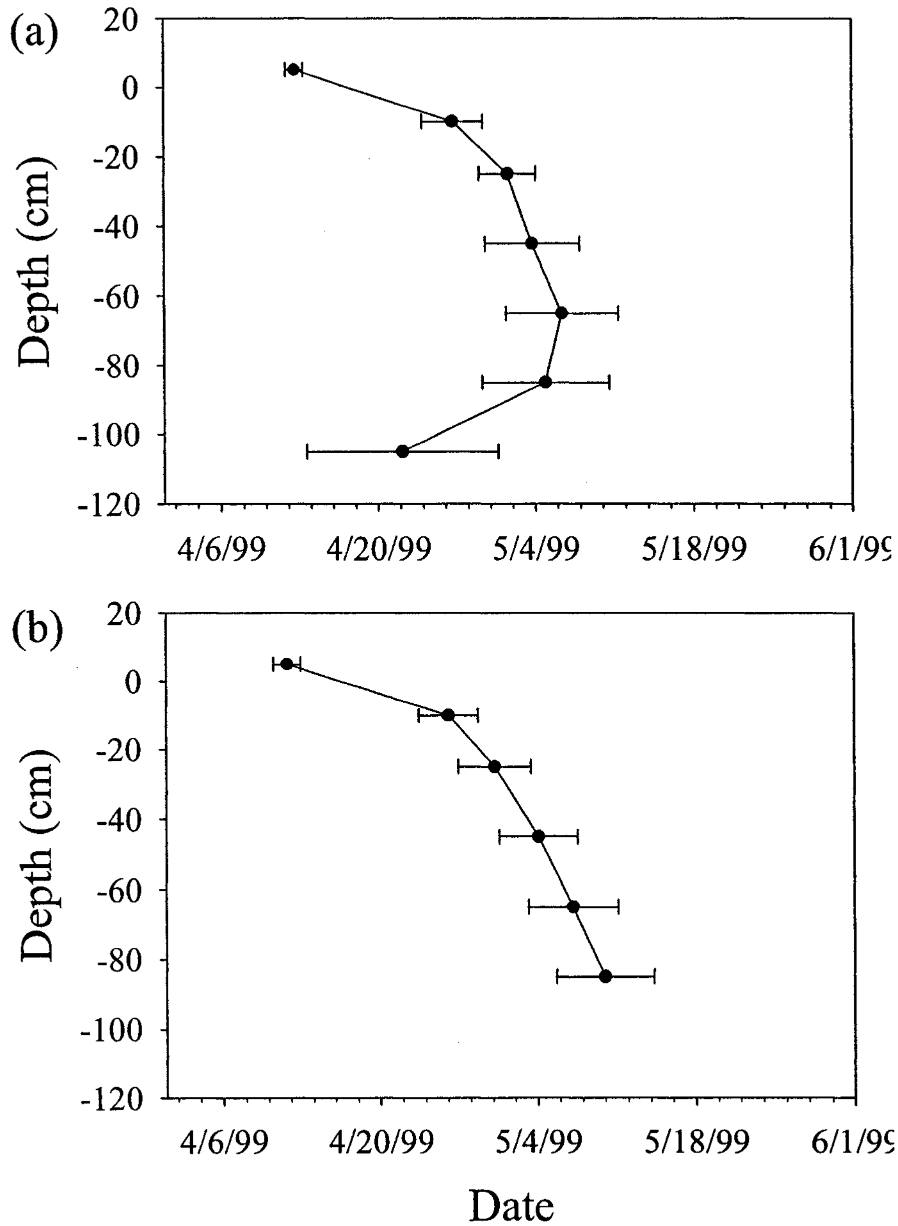



Figure 38: Average and standard error of soil thaw date by increment in a) forested and b) harvested sites

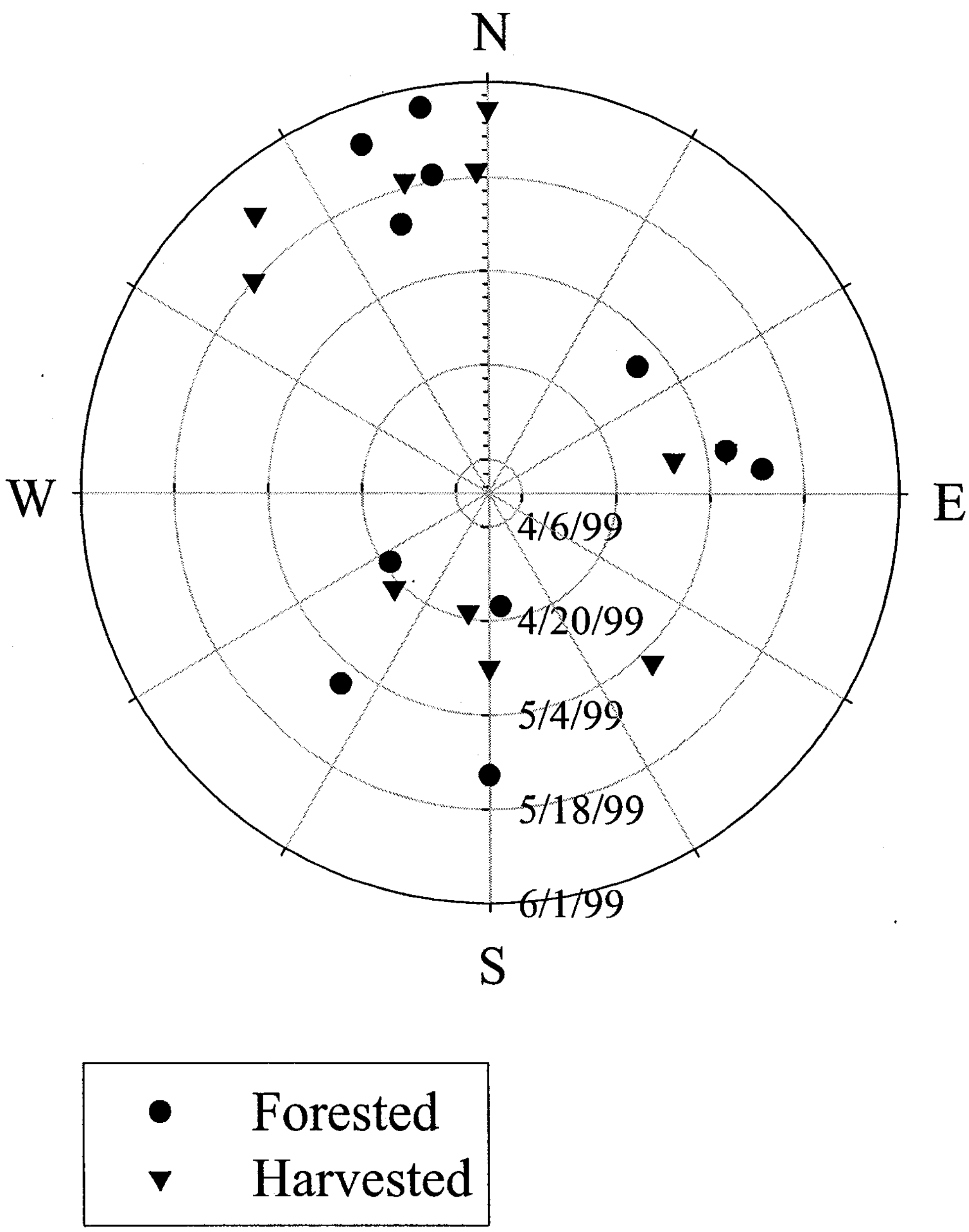



Figure 39: Date of complete soil profile thaw as influenced by aspect and treatment

# Chapter 7

## Synthesis

How does the work described in these previous chapters help us to understand phosphorus movement in Boreal Plain watersheds? Anderson and Burt (1990, p 367) noted that one of the most important areas of hydrology needing research was "the identification of hydrological pathways for subsurface flow." They might have added "and solute transport" because modeling the movement of dissolved substances, particularly reactive ones like phosphorus, is considerably harder than the movement of water alone. High spatial variability of soil properties challenges simulation, but nowhere is the database of these properties sparser than in remote wilderness areas like northern Canada.

The Boreal Plain Ecozone is a 650,000 $km^2$ area of western Canada characterized by sedimentary geology, low-relief physiography, forest vegetation, continental climate, and seasonally frozen soils (Ecological Monitoring and Assessment Network 1996). Pleistocene glaciation created new landforms including till deposits, glacial-lacustrine sediments, outwash and eolian landforms (Ecological Monitoring and Assessment Network 1996). Soils are relatively young and unweathered (Pawluk 1961; Pawluk and Lindsay 1964), or developed on organic material in the case of peatlands. Luvisols are typically associated with the extensive till deposits of the Ecozone (Pawluk and Bayrock 1969).

Because of topographic variation and subsequent change in soil moisture regime, a common catenary sequence includes Luvisols, Gleysols, and Organics (Agriculture Canada Expert Committee on Soil Survey 1987), arranged along a drainage gradient (Strong and Leggat 1992; Beckingham *et al.* 1996). Luvisols, the dominant regional soil (Ecological Monitoring and Assessment Network 1996), occupy better-drained locations, while Gleysols occupy wet areas with higher water tables. Organic soils occupy the wettest locations with shallow water tables. In glacial outwash deposits the Luvisols are replaced by Brunisolic soils (Strong and Leggat 1992; Ecological Monitoring and Assessment Network 1996), while the other members of the catena remain the same. Conceptual models of soil distribution aside, Luvisols dominated the till landscape where my research took place, and comprised the entire soil assemblage on the hillslope where the data for chapters 2 through 5 were obtained.

In the development of hypotheses, I am indebted to the work of Walker and Syers (1976) who suggested that phosphorus loss is most rapid in earlier stages of soil genesis, and of Xiao et al. (1991) who concluded that Gray Luvisols on the Boreal Plain were an example of this rule. Subsequent work on Boreal catenas has shown that losses from Luvisols are associated with gains in downslope Gleysols (Huang and Schoenau 1996). By virtue of organic matter accumulation, Mesisols and Fibrisols of the Organic soil order act as long term phosphorus sinks as long as they continue to aggrade.

Those who adopt a holistic perspective of landscape may question the approach of investigating a flow or solute movement process along a restricted segment of a hillslope catena (e.g., the Luvisolic soil component). Aside from practical considerations (e.g., resources ultimately limiting sample size), I can justify the approach on reductionist grounds. First, a hillslope can be defined as any two-dimensional cross-section of a watershed, extending from drainage divide to streamside. Second, the catenary relationships explored earlier mean that landscape structure can be described by soil types just as a bio-polymer like lignin is conceptualized by the arrangement of its functional groups. An awareness of differences in the processes that affect P movement among major soil groups will allow us to understand where P comes from within the watershed. As an approximation of reality then, I view watersheds as accumulations of individual hillslopes that are in turn described by their soil pattern. Underlying my study of phosphorus movement is a hope that deterministic, process-based landscape simulation can be assisted by empirical studies like mine that break the landscape down into bite-sized chunks.

What then did this study reveal? The very limited amount of subsurface flow during summer (Chapter 2) and snowmelt (Chapter 4) was partially the result of changes in the hydraulic characteristics of upper soil horizons. Interflow generation requires a horizon of much lower hydraulic conductivity, and given that compaction of the Ae by harvest operations probably reduced vertical hydraulic conductivity, these conditions no longer existed. However, low precipitation, only 50% of the long term average, also contributed to reduced interflow because of lower soil water contents, and 1999 runoff in major streams was a mere 9% of the long term average (Environment Canada 2002).

The return period for the storms observed during the study (Appendix 1) was less than 10 years based on calculations developed with reference to a 21 year record from a nearby fire tower. Hence, while on average interflow probably occurs at rates higher than I observed, events are still likely infrequent. The limitation of forecasting interflow from rainfall return periods is that it takes many years of observation to develop the relationship between the two variables, especially when antecedent soil moisture content strongly influences the outcome.

I examined soil hydraulic properties from several perspectives through measurements of infiltration, hydraulic conductivity, bulk density, texture, bromide movement, soil moisture content, and the presence of soil frost, all in the quest to obtain evidence to evaluate the hypothesis of lateral flow. I found that compaction resulting from winter logging conditions, not normally expected to occur (Chapter 3), explained the unexpectedly low Ae horizon saturated hydraulic conductivity and high bulk density the year following and that infiltration rates were still significantly lower in harvested areas three years later. The potential for soil compaction, especially under summer harvest conditions, means that reduced infiltration capacities will increase the probability of overland flow that in turn will transport more phosphorus downslope. I also found support for the hypothesis of anisotropy within the Ae horizon, namely greater lateral hydraulic conductivity than vertical because of soil structure. Because hydraulic conductivity measurements are influenced by the volume of the sample, I used both Uhland cores (small volume) and double ring infiltrometers (large volume) to assess the effect of size. Both datasets were lognormally distributed, despite highly similar soil conditions, emphasizing that the attempt to estimate this variable within narrow confidence intervals will be an ever elusive goal (Warrick and Nielsen 1980).

Bromide tracer studies (Chapters 2 and 6) were inconclusive in demonstrating the existence of lateral flow through upper horizons. The inference that disappearance of bromide over time from the plots was the result of lateral flow was confounded in particular by possible vegetation uptake and by diffusive rather than convective flow. Given the efforts to eliminate plant growth on these plots, it was unlikely however that much of the bromide was absorbed by plant roots, at least over the first winter period when much of the bromide disappeared. The 2.5-cm rain event that occurred during tracer application at transect one in September 1997 "forced" the bromide plume at this plot downward into the Bt where it remained during the next two years of study. It is likely that the rain during application meant the bromide was carried downward in larger soil pores before it could diffuse into small soil pores of lower hydraulic activity (Tindall *et al.* 1999). Forest harvest increased energy inputs and made soils slightly wetter (Chapter 6). Small lateral fluxes will be important for movement of phosphorus and other dissolved constituents downslope towards areas of Gleysolic and Organic soils (muskegs), if not directly into streams and lakes should these lie adjacent.

Water and solute movement during snowmelt has received limited attention in areas where soils are seasonally frozen to a considerable depth as on the Boreal Plain. Logistical challenges alone make studies under such conditions an achievement in themselves. As expected, soil frost played a big part in water flow and phosphorus movement. I found that with dry antecedent soil moisture conditions, snowmelt infiltrated frozen soil, as shown by the rapid response of time domain reflectrometry probes (Figure 20). Most subsurface flow occurred when soil thawed, implying that thaw allowed soil moisture content to increase to saturation in parts of the hillslope (Figures 19c-e, 21c, 24a). Under wetter soil conditions, one would expect more runoff due to lower infiltration (Granger *et al.* 1984). The location of lateral flow in frozen Gray Luvisolic profiles would also change under wet antecedent conditions, with more subsurface flow through the LFH horizon than the Ae horizon. Under such conditions, delayed mineral soil melting (Figure 38) would mean that saturated flow through the LFH could persist for several weeks, depending on aspect. Conditions for leaching of phosphorus from the LFH horizon are likely to reach a maximum during snowmelt under high antecedent soil moisture conditions. Finally, a delay between snowmelt and soil thaw means the runoff response of the watershed will be prolonged well beyond the disappearance of snow.

Although I did not set out to address the question of whether water moved preferentially or by displacement flow, results provided evidence of both. The virtually identical hourly moisture content trends observed (Figure 10) at three widely separated points in both Ae and Bt horizons are examples of how the vertical fluxes at relatively distant points were similar (displacement flow). Results from *in situ* samples showed that preferential flow through larger soil pores affected phosphorus concentrations quite dramatically, at least over short distances (Tables 14 and 16). In these cases, LFH horizon effluents appeared to be undersaturated whereas Ae horizon effluents were usually supersaturated with phosphorus. It appears that flow velocities were too high for desorption reactions to reach equilibrium in the LFH or for adsorption reactions to reach equilibrium in the Ae.

While my work on hydrologic properties of Gray Luvisolic soil profiles has provided some support for the lateral movement of water through upper horizons, it surprisingly found little support for the hypothesis that harvest would increase downslope flux. The work on phosphorus was driven by an interest in exploring the relationship between the mobility of a reactive solute and the mineralogical and chemical properties encountered along that hydrologic flowpath. To summarize Chapter 5, most readily available or soluble phosphorus was in the LFH horizon, where it was predicted to exist in solution largely as orthophosphate or as organic acid-phosphate complexes in equilibrium with several common metal-phosphate solid phases, including calcium-phosphates and an iron-phosphate. Inorganic phosphorus concentrations decreased 3 to 4 orders of magnitude between the LFH and the groundwater zone. Phosphorus retention (adsorption) properties varied with depth, such that the Ae horizon had higher equilibrium phosphorus concentration than lower horizons including the groundwater zone. These findings are in agreement with other studies that have found a vertically stratified solution phosphorus concentration (Evans *et al.* 2000). However, my study has found dissolved phosphorus concentration more than one order of magnitude higher than reported for other natural systems. Soil chemical theory, particularly of solid-phase control of phosphorus solubility, worked well for explaining observed phosphorus distribution. Forest vegetation was found to play a different role in phosphorus movement than expected. Downward organic phosphorus transport with fulvic acid has already been recognized in the region (Schoenau and Bettany 1987). High inorganic phosphorus (phosphate) concentrations observed in my samples and the strong relationship between dissolved organic carbon and phosphorus were explained as complexation reactions between phosphate and organic acids produced by litter decomposition. The dry weather during the study period may also be partially responsible for the observed high solution inorganic phosphorus concentration through reduced leaching. Forest harvest unexpectedly reduced phosphorus availability (Chapter 6). My results were obtained primarily from mixedwood or aspen-dominated sites, and I can only speculate on differences between white spruce and aspen forests. Aspen vegetation in particular cycles base cations more rapidly than spruce and may facilitate soil solutions in which secondary calcium-phosphates can form. An increase in pH in a calcium-phosphate system reduces phosphorus availability, whereas in an aluminum- or iron-phosphate situation, it increases it (Lindsay 1979).

The related TROLS study conducted on Boreal Plain lakes at the same time as my own research found that the effect of harvest on water quality depended on precipitation (Prepas *et al.* 2001). Total phosphorus concentration in lakes increased during the first post-treatment year when precipitation was unusually high, but did not increase during the second post-treatment year when weather was uncharacteristically dry. The results of my research suggest that the net effect of harvest will depend on the extent that harvest increases runoff and reduces P concentration in moving water. Harvest resulted in less interflow due to changed hydraulic properties, and did not result in surface runoff during snowmelt as surface horizons still had high infiltration capacities even during frozen conditions. Furthermore, because of reduced P concentrations at these sites, harvest probably decreased the normal downslope flux of P to lower slope positions. A change to higher flowpaths though, such as through the LFH horizon that might happen in frozen, moist soils with low air filled porosity, would lead to increased flux of P downslope. Gray Luvisols have high solution SRP in the litter layer, that, if mobilized, could yield large fluxes of phosphorus to positions lower in the landscape. I found mean soluble reactive phosphorus concentrations in LFH horizon pore-water that were 70 to 3000 times greater than those reported for soil solutions of a range of Boreal Plain wetland types (Vitt *et al.* 1995). While fluxes of water from Gray Luvisolic soils are likely to be lower than from other soils in the watershed, particularly Gleysols and Organics, their contribution to the flux of phosphorus within the landscape will be appreciable because of higher solution phosphorus concentration. Therefore, their contribution to phosphorus flux within watersheds should always be considered.

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*Appendix 1: Storm Frequency Estimation*

Storm frequency was analysed using long-term precipitation records from Rock Island Fire Tower located 20 km southeast of the study site, where precipitation was recorded twice-daily typically from April to September by staff of the Alberta Forest Service. Rainfall records for the period 1977 to 1997 were used to develop a partial duration series for storm events greater than 10 mm in magnitude and for durations from 1 to 5 days. Return periods for each duration were calculated assuming a Log Pearson 3 distribution (Haan 1977). Results were approximately linear and allowed estimation of the return periods of storm events for 1998 and 1999.

One would expect a relationship between storm magnitude and the amount of subsequent subsurface flow. I estimated the frequency of storms that had produced measurable subsurface flow with measurements of depth and duration recorded at Rock Island and using storm frequency estimates developed from the 1977-97 database from that same location (Figure 40). Five runoff events occurred in 1998 and only one in 1999, excluding the snowmelt period. When 1998 and 1999 runoff events are ranked in terms of volume of subsurface flow, they somewhat match the rank of the associated storm event at Rock Island tower 20 km distant (Table 24). The July 25-27, 1998 flow event was the largest; the associated storm event at Rock Island had a return period of between 5 and 10 years. Next largest was the flow event of 1999, generated by a storm of about a 2-year return period. Third in terms of runoff was the flow event of June 28, 1998, that had an associated return period of between 2 and 5 years. Remaining flow events had less runoff and were associated with storm events with return periods of less than 2 years. Interflow was also observed on July 19, 1997 in a soil pit the day after a storm event with an estimated return period of between 5 and 10 years. Thus, a relationship between storm magnitude and the magnitude of interflow is apparent, as expected.

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Table 24: Comparison of storm frequency at Rock Island and runoff at study hillslope

| | Storm Characteristics | | | |
|---|---|---|---|---|
| *Date* | Duration (days) | Precipitation (mm) | Return Period (y) | Runoff volume (ml) from all flow collectors |
| *July 25, 1998* | 2 | 36.8 | $5<x<10$ | 252 |
| June 28, 1998 | 2 | 23.4 | $2<x<5$ | 89 |
| August 21, 1999 | 1 | 14.7 | $x\sim2$ | 150 |
| July 11, 1998 | 1 | 12.4 | $x<2$ | 30 |
| August 4, 1998 | 1 | 10.3 | $x<2$ | 14 |
| September 25, 1998 | 1 | 7.1 | na | 15 |

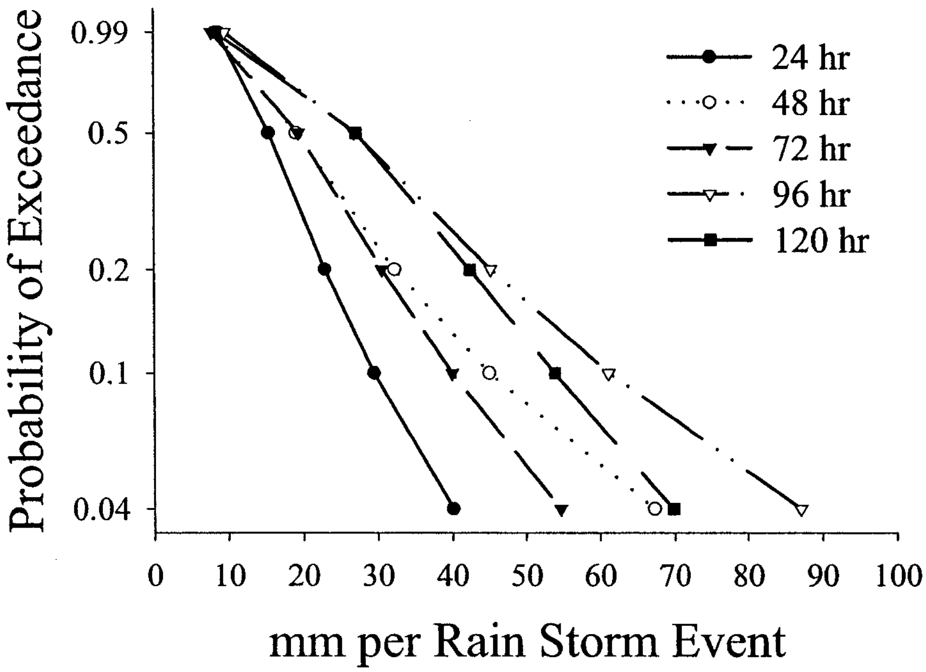


Figure 40: Storm frequency estimates for Rock Island Fire Tower

*Appendix 2. Explanation of Terms*

| | |
|---|---|
| A (horizon) | Designation for the first mineral soil horizon in the Canadian System of Soil Classification (Agriculture Canada Expert Committee on Soil Survey 1987). |
| Ae (horizon) | Designation used in the Canadian System of Soil Classification; denotes clay removal as the dominant process in the A horizon. |
| Alfisol | Category in the uppermost level of the US Soil Taxonomic system; soil with high clay content in the subsoil and high base saturation status. Generally equivalent to Luvisol of the Canadian System of Soil Classification. |
| Anisotropic | Case where the hydraulic conductivity at a point in the soil differs with direction |
| BC (horizon) | Transitional horizon, having properties of both the horizons immediately above (B) and below (C) |
| Boreal Plain | Category in the Ecological landscape classification system that denotes cold, continental climate (cold winters, warm, moist summers), forest vegetation, and sedimentary geology. Unit occurs in northwestern North America east of the Rocky Mountains, west of the Canadian Shield, south of the zone of permafrost occurrence, and north of the prairie grassland ecosystem. |
| Bt (horizon) | Horizon of higher clay content relative to horizons above. |
| Bulk Density | Weight of soil particles per unit volume after drying to 105 °C. |
| Ck (horizon) | Denotes a parent material (unweathered) horizon that contains calcium or magnesium carbonate. |
| Clay | Particle falling within the size class of 0 to 0.002 mm. |
| Isotropic | Case where the hydraulic conductivity at a point in the soil is equal in all directions. |
| Interplanar voids | Voids between soil peds. |
| LFH (horizon) | Horizon designation that describes litter accumulation at the forest floor. |
| Luvisolic; Luvisol | Soil category at the uppermost level of the Canadian System of soil classification; denotes physical translocation of silicate clay from the Ae to the Bt horizon and high base saturation. |
| Ped | An aggregation of soil particles that tends to retain a particular shape and size. Each horizon has a characteristic kind of ped that is a function of the soils structure. |
| Pedon | Smallest conceptual genetic unit of a soil; *see profile.* |
| Percolation | Downward movement of soil water |
| Platy (structure) | Aggregates breaking along horizontal cleavage planes. |
| Porosity | Proportion of the soil bulk volume occupied by air or water; bulk volume less solid volume. |

| | |
|---|---|
| Profile | Two-dimensional crosssection of a pedon; commonly viewed with soil cores or soil pit faces. |
| Sand | Soil particle greater in size than 0.050 mm and less than or equal to 2.000 mm. |
| Silt | Soil particle greater in size than 0.002 mm and less than or equal to 0.050 mm. |
| Structure | Aggregation of soil particles into collective units (peds) that can be described by size and geometry. |
| Subangular blocky structure | Aggregation of soil particles that break into geometric shapes with greater than 6 planar sides. |
| Subsoil | Collective term for soil horizons below the A horizon and above the parent material horizons. |
| Texture | Proportion by weight of sand, silt, and clay. |
| Ultisolic; Ultisol | Category in the upper most level of the US Soil Taxonomic system; soil with high clay content in subsoil, low base saturation status. |

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