Abstract. In order to demonstrate the spatial and temporal variations in meltwater chemistry at both the lysimeter (0.25 m²) and basin scale, field measurements of snowmelt were conducted in northern Canada. These observations show that microscale variations in chemistry at both the lysimeter and basin scale, field measurements of solute concentrations were largest in areas with low flow, while the largest mass flux occurred in the areas with highest flow. The observed variations in both concentration and mass flux can be quantitatively described by the relationships described by Hibberd [1984]. The field measurements clearly demonstrate that in order to estimate the average meltwater chemistry, it is necessary to sample the flow field at a scale similar to that required to average the lateral variations in meltwater volume. Variations in meltwater runoff chemistry also occur at the basin scale due to changes in snowcover depth and the resulting differences in the timing of meltwater release. For example, at this site, meltwater release occurs up to a week earlier from the shallow snow covers than for the deeper snow covers. It would be expected that this asynchronous meltwater runoff would result in a smoothing of the ionic pulse at the basin scale, with lower peak values and a more gradual decline in concentration when compared with meltwater at a point.

1. Introduction

Chemical species accumulate in northern and alpine snowpacks for up to 8 months of the year. The melt of these snow covers results in high total loads and peak concentrations of the chemical species in the meltwater released from the snowpack [Jones, 1987; Tranter et al., 1987]. Typically, 80% of the total loading is released in the first 20–30% of the melt period [Bramblecombe et al., 1986]. This ionic pulse can impact micro-biological growth within the snowpack [Hoham et al., 1993], result in the delivery of nutrients to the soil at the initiation of the growing period, and have detrimental impacts on forest ecosystems [Abrahamsen et al., 1977; Stroo and Alexander, 1986]. In areas where the flow pathways transmit meltwater directly to the stream channel with little interaction with the soil [Ferrier et al., 1989], the ionic pulse may pass directly to the stream channel, with a resulting rapid rise in the ionic concentration of the stream water. In regions where the soil provides little buffering, there may be a temporary acidification of streams and lakes [Wigington et al., 1990; Tranter et al., 1988], resulting in stress conditions for the aquatic biota and, in some cases, fish mortality [Gunn and Keller, 1984; Driscoll et al., 1980].

The high concentrations in the early portions of melt (fractionation) [Tranter, 1991] and the fact that certain chemical species are released prior to others (preferential elution) [Tranter, 1991] have been explained by the microscale variations in ion location on and within snow particles due to chemical exclusion during ice crystal growth [Hewitt et al., 1991]. Although the ionic pulse always occurs in the early phases of melt, previous studies [e.g., Tranter, 1991] have noted considerable variations in the magnitude of the pulse. For example, the magnitude of fractionation, expressed as a concentration factor CF (equal to C/C₀, the ratio of solute concentration in the flow to the initial snowpack solute concentration), typically varies from 2 to 10. Some of the factors used to explain these variations include melt rate [Tranter, 1991], location of the solute in the snowpack [Bales et al., 1989], freeze-thaw [Colbeck, 1981], and snow depth [Davies et al., 1987]. However, the spatial and temporal variations in CF have not been well documented. For example, heterogeneous flow paths (P. M. Foster [1978 report referenced by Tranter [1991]), also Harrington et al. [1996]) would be expected to exert significant variations in CF at the microscale, while differences in timing of wetting front movement through the snowpack would be expected to control the timing of the ionic pulse at the stream basin scale.

With these deficiencies in mind, the objective of this paper is to outline the spatial and temporal variations in meltwater flow volume and chemistry within natural snowpacks at both microscale and basin scale. Specifically, we will investigate the impact of flow variability on the CF and mass flux of meltwater and finally consider the potential role that the spatial variations in snow depth play in controlling the timing of meltwater runoff chemistry.

2. Methods and Study Area

Field work was carried out during May and June 1992 near Trail Valley Creek, located approximately 50 km north-northeast of Inuvik, Northwest Territories (NWT), and 80 km south of the Beaufort Sea. Snow cover typically begins to develop in this area in late September or early October, melt begins in mid to late May, and the snow cover is completely removed by mid-June. The mean daily January temperature at Inuvik is –30°C, and the mean annual precipitation is 266 mm, of which 57% falls as snow. Trail Valley Creek is within the zone of continuous permafrost and consists of an upland tun-
The transport of water and solutes through this Arctic
from 70 to 140 m above sea level. Low shrub tundra vegetation
depth [Pomeroy et al., 1993].

Variations in snow water equivalent were described for the
63-km² Trail Valley Creek basin by Pomeroy et al. [1993, 1997].

Tests demonstrated that filter size did not affect the results.

A 0.25-m² lysimeter, divided into 16 separate compartments
each 0.0156 m² in area, was installed 35 cm above the base of
this snowpack. Each compartment was lined with a plastic
liner, which was rinsed repeatedly with deionized water prior
to installation. The lysimeter was installed into a snow pit wall
as the pack approached isothermal conditions, using the meth-
odology described by Marsh and Woo [1985]. This lysimeter
design and installation method ensured that there was no con-
tamination of the collected water by water that had contacted
the soil. Meltwater from each compartment was collected in
individual containers, and the cumulative flow volume was
measured daily at the time of minimum flow. This typically
occurred in the early morning when flow was near zero. This
sampling scheme ensured that each sample contained flow
originating over a single melt day. Meltwater from four com-
partments was saved each day for subsequent chemical analy-
sis. These samples included the compartments with the lowest
and highest flows, while two other samples were saved to rep-
resent compartments with flows between these extremes. Since
the compartments with the highest and lowest flow varied from
day to day, the chemistry samples were often obtained from
different compartments on subsequent days. Snow physical
properties, including the location of the wetting front during
the melt period, and premelt snow chemistry were measured in
a second snow pit located approximately 2 m away from the
lysimeter pit. Additional data from a study conducted near
Resolute Bay, NWT, in June 1990 will also be used in the
present paper. Details of this study site are provided by Marsh
and Woo [1985].

All snow and meltwater samples were collected, prepared,
and analyzed as follows. Snow pits were dug with Teflon-
coated shovels and scraped clean with scoops that had been
soaked and rinsed with deionized water. Snow was sampled
from distinct snow strata and snow density was measured from
each layer. Snow samples were collected in clean polyethylene
resealable bags, while lysimeter samples were collected in high-
density polyethylene (HDPE) bottles that had been washed in
nitric acid and then rinsed several times with deionized water.
Melted snow samples and lysimeter meltwater samples were
filtered through either 0.2-µm or 0.4-µm Nucleopore filters.
Tests demonstrated that filter size did not affect the results.
The pH was measured with an Orion 720A meter with a Ross
Combination pH electrode before and after filtering. There
were small increases in pH after filtering; the before-filtering
values are used here. All samples were then stored in HDPE
bottles that had been prewashed as noted above, and then
shipped from Inuvik to Saskatoon for analysis.

Anion analyses (Cl⁻, NO₃⁻, SO₄²⁻) were conducted on a
Dionex 210iP ion chromatograph with a 100-µl injection loop,
0.75 m NaHCO₃/1.5 m Na₂CO₃ eluent, and suppressed con-
ductivity detection. Samples were introduced by Technicon
 sampler into the loop after mixing nine parts sample to one part
concentrated eluent (10x the working eluent concentra-
tion). A Dionex 4270 Integrator measured peak area, and a
Linear 100 recorder measured peak height of Dionex concen-
tration tracings. Cation analyses (Ca²⁺, Mg²⁺, K⁺, Na⁺) were
performed using flame atomic absorption spectrophotometry
on Perkin Elmer AAS models 460, 2380, and 5000. Flame
mixtures for Na and K were Air:H₂, for Mg were Air:Acetylene
and for Ca were N₂O:Acetylene. For Mg analyses, 1000 μg/ml
Na was added to samples and standards, and for Ca analyses,
2000 μg/ml K was added. Numerous blanks on scoops, bags,
bottles, filters, and analytical equipment were consistently be-
low detection limits, indicating that no measurable contamina-
tion of samples occurred. The level of precision and limits of
detection of the AAS Dionex is normally at least an order of
magnitude less than the levels measured in the meltwater and
snow samples.

A concentration factor CF was calculated for the lysimeter
runoff samples from CF = C/Cₒ, where C is the concentra-
tion in the meltwater outflow and Cₒ is the weighted mean concen-
tration in the premelt snow cover. Ion charge balance (cations
– anions) were performed on each snow stratum in the
premelt snowpack and for the meltwater discharge from the
lysimeters as a quality control index. Charge balance residuals
were typically small compared with the total anions plus cat-
ton concentrations. For example, early in the melt period when
the concentrations were high (total anions and cations >120 μeq/l),
the charge balance residuals were typically ∼10–15 μeq/l, while
later in the melt period, when the concentrations were
low (total anions plus cations over 30 μeq/l) the charge balance
residual was typically <5 μeq/l. These low residuals suggest
that most major ions were measured.

3. Results

3.1. Premelt Snow Cover

The premelt snowpack, sampled on May 19, 1992, near the
lysimeter site, was 125 cm in depth, had a mean density of 352
kg/m³, and a total snow water equivalent (SWE) of 440 mm.
The thickness, SWE, anion and cation concentrations of each
layer are given in Table 1. Although there were large variations
in H⁺, Cl⁻, SO₄²⁻, Ca²⁺, Mg²⁺, Na⁺, and K⁺ concentrations
between layers, there were no consistent trends from top
to bottom of the pack. NO₃⁻, however, was considerably different.
The top layers of the pack had NO₃⁻ concentrations 1–2 orders
of magnitude larger than those in the bottom layers of the
pack. Pomeroy et al. [1993] and Pomeroy and Jones [1996]
suggest that low concentrations of NO₃⁻ in the bottom of the
pack are due to depletion during temperature gradient meta-
morphosis over the winter. Because of the variations in SWE
between layers, a weighted mean concentration over the ver-
tical profile was calculated for each ion (Table 1).

3.2. Microscale Spatial Variations in Meltwater Chemistry

3.2.1. Flow variability. Heterogeneous flow paths are ini-
tiated when meltwater first enters a snow cover and the wetting
Table 1. Premelt Snow Pit Data at the Multicompartent Lysimeter Site, May 19, 1992 (Day 140)

<table>
<thead>
<tr>
<th>Description</th>
<th>Layer</th>
<th>Thickness, mm</th>
<th>SWE, mm</th>
<th>Ion Concentration, μg/L</th>
</tr>
</thead>
<tbody>
<tr>
<td>Top layer: new snow</td>
<td>9</td>
<td>40</td>
<td>14</td>
<td>13.2 33.3 5.2 10.4 5.0 10.0 23.0 1.0</td>
</tr>
<tr>
<td>Wind crust</td>
<td>8</td>
<td>140</td>
<td>0</td>
<td>12.3 14.8 3.3 5.6 1.5 4.2 9.1 0.5</td>
</tr>
<tr>
<td></td>
<td>7</td>
<td>140</td>
<td>60</td>
<td>11.5 8.4 3.1 4.8 0.3 2.5 5.2 0.8</td>
</tr>
<tr>
<td>Ice layer</td>
<td>5</td>
<td>10</td>
<td>7</td>
<td>13.8 26.7 6.2 7.5 4.0 6.7 17.8 1.0</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>50</td>
<td>17</td>
<td>2.0 49.9 0.05 9.2 11.0 18.3 37.8 6.4</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>390</td>
<td>138</td>
<td>4.6 22.9 0.05 8.1 5.5 7.5 19.1 1.0</td>
</tr>
<tr>
<td>Bottom layer</td>
<td>2</td>
<td>200</td>
<td>54</td>
<td>1.0 24.9 0.82 6.9 14.0 9.2 20.0 0.5</td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td>1250</td>
<td>440</td>
<td>Mean 8.3 25.8 2.7 7.5 5.9 8.3 18.9 1.6</td>
</tr>
<tr>
<td>Mean Weighted mean</td>
<td></td>
<td>5.5</td>
<td>29.6</td>
<td>1.4 7.5 7.9 10.3 22.3 2.5</td>
</tr>
</tbody>
</table>

The snow water equivalent (SWE) for the ice layer is estimated assuming a density of 700 kg/m³. Also shown are the mean concentrations and the mean concentrations weighted by SWE for each snow layer.

Front develops flow fingers at its leading edge [Marsh, 1991]. These flow fingers develop as a result of flow instabilities at the leading edge of the wetting front [e.g., Parlange, 1974] and structural variations in the snow. The permeability of the snow in the flow finger increases as a result of more rapid grain growth [Colbeck, 1976; Marsh, 1987] due to the higher water content in the flow path as compared with the surrounding "nonflow path" zone. These higher-permeability pathways may remain for much of the snowmelt period. The pathways, in combination with horizontal ice layers, which decrease flow in certain areas and increase it in others, result in spatial variations in the downward meltwater flux. Few measurements have been made of this flow variability, but it has been suggested that the surface area required to average the spatial variations in flow is of the order of $z^2$, where $z$ is the snow depth [Male and Gray, 1981]. Field measurements of the variability in vertical flow in shallow Arctic snowpacks ($z$ from 1 to 1.5 m), has shown that flow is relatively uniform over areas of approximately 1 m², while within 0.02-m² areas flow varies from 0.2 to 3 times the mean flow over a larger area [Marsh and Woo, 1985]. In deeper snow of the Sierra Nevada Mountains ($z$ up to 2.4 m), Kattelmann [1989] found that flow could be nonuniform over areas of several square metres.

Figure 1 shows the total daily flow from each of the 16 compartments in the lysimeter, with flow expressed as a fraction of the mean daily flow for all compartments, for the period from day 149 to day 164. From day 149 to 155, daily maximum air temperature ranged from 5.9°C to 9.6°C, and daily...
variability had decreased substantially, with compartment CF values). In addition, Figure 2 demonstrates the effect of ions are enriched more than others, as indicated by different variations in flow are similar to those reported from other Arctic sites.

Flow resumed again on account of subzero air temperatures. Flow resumed again on day 162 and temperatures remained above zero for the remainder of the melt period. This figure clearly shows fractionation through homogeneous snow developed by Colbeck [1985].

Figure 2. Concentration factors CF for the compartments with the highest (open symbols) and lowest (solid symbols) flows. Over the time period plotted, ~80% of the snowpack was removed.

The maximum CFs for most of the ions are similar to those reported by other studies [Tranter, 1991]; however, that for NO₃⁻ was significantly higher. This is probably due to the fact that the premelt concentration of NO₃⁻ is much higher in the upper portions of the snow pack. As reported by Bales et al. [1989], this irregular, layered distribution results in higher CFs than if the solute is evenly spread through the pack. Using the measured meltwater concentrations and the premelt snowpack concentration in the upper portion of the pack for reference, results in a calculated CF of between 2 and 3 for NO₃⁻.

Although the high-flow areas showed low CF, and vice versa, flow rates still determined the relative daily ionic mass flux. Figure 3 shows that the highest mass fluxes occurred from the high-flow areas, showing that high flow volumes overcome the lower concentrations in determining the location of highest ionic mass flux. These data demonstrate that the CF relationship with flow volume is not simply a dilution process. If it was, then each flow zone would have different concentrations, but all would have similar ionic mass flux. Instead, with increasing flow volume, more ionic mass is removed from the snow particles but it is not in direct proportion to flow. As a result, the mass flux increases and the concentration decreases.

Preferential elution is demonstrated in Figure 4, which shows changes in ionic ratios with time. The decreasing SO₄²⁻/Cl⁻ and NO₃⁻/Cl⁻ ratios suggest that SO₄²⁻ and NO₃⁻ elute earlier than the Cl⁻, while the nearly constant SO₄²⁻/NO₃⁻ ratios imply that these ions do not elute differentially. It is interesting to note that only the NO₃⁻/Cl⁻ ratios vary considerably with flow volume, with the low-flow compartment having higher ratios early in the melt period.

3.2.3. Discussion. Numerous studies [e.g., Harrington et al., 1996] have provided at least a qualitative understanding of the role of flow volume on the processes responsible for fractionation and preferential elution and therefore solute concentration in meltwater; however, there have been few modelling studies [Hibberd, 1984; Bales, 1991; Harrington and Bales, 1998]. Hibberd [1984] presented a simple model for predicting meltwater chemistry that made the following assumptions: (1) Heat flux is uniform at the surface, and there is no solute input from the surface. (2) Solutes are evenly distributed through the snowpack. (3) Solutes are held in the irreducible water content. (4) Solutes are transferred from the immobile water to the mobile water with the arrival of the first meltwater wavefront and then dispersed within the meltwater by hydrodynamic dispersion (molecular diffusion and mechanical dispersion). Hibberd [1984] used the one-dimensional theory of water percolation through homogeneous snow developed by Colbeck...
Figure 3. Mass flux for the compartments with the highest (open symbols) and lowest (solid symbols) flow. Note change in scale for the bottom four graphs.

Figure 4. Anion concentration ratios for each of the low-flow (solid symbols) and high-flow (open symbols) compartments in the multicompartment lysimeters. Changes in ratios with time suggest that the ions elute at different rates.
and concentration can be explained by the variations in flow velocity, the transfer of solutes from irreducible water (high solute concentration), to meltwater (lower concentration), and the hydrodynamic dispersion of solutes within the meltwater.

There are many factors influencing CF and ion mass flux in snow that are not incorporated into the Hibberd [1984] model or the modification by Bales [1991]. One example are ice layers that grow at premelt stratigraphic layers within the melting snowpack [Marsh and Woo, 1984; Kattelmann, 1995]. Figure 6 shows changes in $NO_3^-$ concentration for both the snow and an individual ice layer within a melting snow cover in the Resolute Bay area. Prior to the start of melt, the $NO_3^-$ concentration was 2.8 μeq/L, and during the melt period the $NO_3^-$ concentration gradually declined to a minimum value of 0.6 μeq/L (Figure 6). However, an ice layer that formed within the snowpack on June 16, 1990 had an initial concentration significantly higher than the snow (6.4 μeq/L). The concentration then gradually declined, reaching a value similar to that of the snow by June 22. This example suggests that for cold snow covers, ice layers may temporarily store solutes, thus limiting the peak CF of meltwater runoff and increasing the CF later in the melt pe-

**Figure 5.** Relationship between CF and mass flux, and water saturation (and the equivalent water flux) using data from Hibberd [1984, Figure 6]. Snowpack properties are provided in the figure, where $S_i$ is the irreducible liquid water content of the snow, $n$ is a parameter in the meltwater-liquid water relationship, and $k_s$ is the saturated permeability.

**Figure 6.** Changes in $NO_3^-$ concentration for snow and an ice layer (Resolute Bay, NWT) during the melt period in June 1990. This ice layer did not exist prior to the beginning of melt but was formed by the refreezing of meltwater within the snowpack (see Marsh [1991] for description).
3.3. Basin Scale Variations in Snowmelt Chemistry

High-concentration meltwaters are not always passed directly into the stream for a number of reasons [Wolford et al., 1996], including the impact of the various hydrologic pathways that the meltwater takes to the stream channel. For example, the early release of ions means greater likelihood that the meltwater will infiltrate the frozen soils rather than run off as surface flow, resulting in a terrestrial rather than an aquatic input. Another mitigating factor that is often not considered is the occurrence of spatial variations in the timing of meltwater release. For example, it is quite common for the release of snow meltwater to be nonsynchronous over a basin. This occurrence can be due to spatial variations in melt rate, as would occur at different elevations, slopes, and aspects in mountainous areas, or due to snow-depth-induced differences in the meltwater release.

Pomeroy et al. [1993] described variations in the snow depth in Trail Valley Creek, with the tundra areas having a mean depth of 45 cm, while the shrub tundra and drift areas had mean depths of 100 cm and 185 cm, respectively, at the end of winter. Even though the drift areas cover only 8% of the catchment, approximately one third of the total snow stored in the basin is found in these areas [Marsh and Pomeroy, 1996]. It would be expected that these large variations in snowcover would significantly affect the timing, magnitude and chemistry of meltwater released to the stream channel due to differences in the rate of wetting front advance through the snow.

With the start of snowmelt, meltwater infiltrates into the snowpack, dividing the pack into an upper zone which is wet and isothermal at 0°C, and a lower zone which is dry and below 0°C. The wetting front dividing these two zones is generally not uniform [Marsh and Woo, 1985], but instead is characterized by small flow fingers that cover approximately 20% of the horizontal area [Marsh and Woo, 1985]. The result may be conceptualized as a two-component wetting front, with a finger and background front. The finger wetting front represents the deepest penetration of liquid water into the snowpack, and the background wetting front denotes the depth above which all
areas (-70% of the basin) were contributing water, while the finger fronts for the deeper shrub and tundra sites delayed by the early portions of the melt period, only the shallow tundra site at Trail Valley Creek. These results demonstrated that the weighting the CF for each terrain type by the area of the shallow tundra snowpacks released meltwater first, with the drift areas not contributing meltwater. Likewise, later in the melt period the tundra snow was completely melted, while the drift areas were still contributing meltwater.

The potential impact of these variations in wetting fronts on the timing of NO$_3^-$ CF curves for each terrain type is shown in Figure 7. In this case the CF curve for each terrain type is assumed to be the same as the mean of the CF curves shown in Figure 2, but with the timing lagged according to the arrival of the predicted wetting front at each of the tundra, shrub, and drift sites. As was suggested by Harrington and Bales [1998] this assumption of similar CF for snowpacks of different depth is reasonable. A basin mean CF (Figure 7) was then calculated by weighting the CF for each terrain type by the area of the terrain type and the fractional snow cover of each terrain type (Figure 7). The result is that the basin CF is initially the same as that for the tundra but then remains slightly higher than the tundra as the shrub and drift begin contributing.

The impact of this asynchronous meltwater pulse on streamflow would be expected to vary greatly depending on the meltwater delivery mechanism, with chemical transformations depending on the duration of contact with the substrate and the substrate material. In areas of the high Arctic, where frozen soil infiltration is low and the meltwater has limited contact with the substrate, it would be expected that the streamflow chemistry would be closely related to that of the meltwater released from the base of the snowpack. In contrast to this, in many subarctic areas, most meltwater runoff follows various subsurface pathways through peat [Quinton, 1997] to the stream channel. The meltwater may then undergo various transformations prior to reaching the stream channel, and the streamwater chemistry may be substantially different from the original meltwater.

4. Conclusions

The field observations reported in this paper provide a direct measurement of the spatial and temporal variations in meltwater chemistry. These observations show that in a typical deep (1.25 m) snowpack, flow early in snowmelt varies from 0.25 to 3 times the mean over small areas, and the variability decreases later in the melt period to between 0.5 and 1.5 times the mean flow. In conjunction with these observed variations in flow volume, it was found that the concentration of solutes within the meltwater also varied greatly. The concentration factor CF was higher in zones of low flow than in zones of high flow, but the ionic mass flux was highest in zones of high flow. The model of Hibbard [1984] can at least qualitatively describe these relationships between CF, mass flux, and flow volume. Over time, the differences in solute concentration between high- and low-flow areas decreased, and after approximately 50% of the observed snowmelt, the high- and low-flow zones had similar CFs, although high-flow areas continued to sustain higher ionic mass fluxes. These observations demonstrate that the CF measured over "large" areas in snowpacks is a result of the mixing of water with a wide range of CF's. These observations clearly demonstrate that the CF measured in lysimeter size is dependent on the scale of measurements. This suggests that at least some of the variability in CF's reported in the literature may be due to differences in lysimeter size in relation to the representative scale of variability for any given snowpack. For example, in a shallow snowpack, a small lysimeter might provide a representative sample, whereas the same size lysimeter in a deep snowpack will not.

Variation in snow cover depth between different terrain types, results in differences in the timing of meltwater release. Therefore the timing of the ionic pulse is asynchronous, with shallow snow areas releasing high CF water 5 days earlier than the deep drift areas. This has important implications for the timing and magnitude of the ionic pulse to the soil and, in some cases, to the stream channels.

Acknowledgments. We would like to thank the Polar Continental Shelf Project and the Aurora Research Centre for there generous logistical support. In addition, we would like to thank Mary Ferguson, Cuyler Onclin, Bill Quinton, Jeannie Patterson, and Ken Supeene for their help in collecting and analyzing the snow and meltwater samples.

References


Hewitt, A. D., J. H. Cragin, and S. C. Colbeck, Effects of crystal metamorphism on the elution of chemical species from snow, in
MARSH AND POMEROY: SNOWMELT RUNOFF CHEMISTRY VARIATIONS

Quinton, W. L., Runoff from hummock covered Arctic tundra hillslopes in the continuous permafrost zone, Ph.D. dissertation, 277 pp., Univ. of Sask., Saskatoon, Canada, 1997.

P. Marsh and J. W. Pomeroy, National Water Research Institute, National Hydrology Research Centre, 11 Innovation Boulevard, Saskatoon, Saskatchewan, Canada S7N 3H5. (marshp@nhrisv.nhrc.sk.ec.gc.ca)

(Received March 2, 1998; revised December 1, 1998; accepted December 2, 1998.)