Characterizing hydrodynamics on boreal landscapes using archived synthetic aperture radar imagery

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Abstract:
Characterizing the spatial and temporal variation in surface hydrological dynamics of large boreal landscapes is vital, since these patterns define the occurrence of key areas of land-to-lake and land-to-atmosphere hydrological and biogeochemical linkages that are critical in the movement of matter and energy at local to global scales. However, monitoring surface hydrological dynamics over large geographic extents and over long periods of time is a challenge for hydrologists, as traditional point measurements are not practical. In this study we used European Remote Sensing satellite radar imagery to monitor the variation in surface hydrological patterns over a 12-year period and to assess the change in the organization of saturated and inundated areas of the landscape. Using the regional Utikuma River drainage basin (2900 km²) as the test area, the analyses of patterns of wetlands indicated that, during dry climatic conditions, wetland sizes were small and disconnected from each other and receiving bodies of water. As climatic conditions changed from dry to mesic, wetland numbers increased but were still disconnected. Very wet climatic conditions were required before the disconnected wetlands coalesced and connected to lakes. During these wet conditions, the response of the lake level at Utikuma Lake was observed to be much higher than under drier conditions. Analyses of individual wetland maps and integrated wetland probability maps have the potential to inform future biogeochemical and ecological investigations and forest management on the Boreal Plain. Copyright © 2007 John Wiley & Sons, Ltd.

INTRODUCTION
Recent acceleration of human activities such as oil, gas and timber extraction on the Boreal Plain of northern Alberta has raised concerns over possible disruptions and/or alterations of existing surface hydrological dynamics and associated impacts on terrestrial and aquatic ecosystems (Prepas et al., 2001; Schindler, 2001; Smith et al., 2003). Surface hydrological features (i.e. saturated and inundated areas) demand monitoring and potentially protection, since they play a significant role in mediating the transfer of water (e.g. Bowling et al., 2003; Lindsay et al., 2005), nutrients (e.g. Creed and Band, 1998; Devito et al., 2000; Evans et al., 2000; Macrae et al., 2005), sediments (e.g. Rustomji and Prosser, 2001; Chaplot et al., 2005), and biota (e.g. Cottenie and De Meester, 2003; Pringle, 2003; Ray et al., 2004) from terrestrial to aquatic systems and in the exchange of gases between terrestrial and atmospheric systems (e.g. Savage et al., 1997). In order to monitor surface hydrological dynamics, techniques are needed that provide hydrological data over large areas (e.g. regional drainage basins) because management decisions are being made at these scales. Ground-based techniques provide point measurements of soil moisture (Munoz-Carpena, 2002) and, therefore, they are not feasible for large-scale mapping. Advances in remote sensing allow hydrologists to move beyond point measurements and to monitor soil moisture and patterns of inundation at broad spatial scales (Toyra et al., 2001; Moran et al., 2004; Western et al., 2004). This paper explores the use of satellite radar imagery in mapping wetland areas, including both saturated and inundated lands.

The dominant processes controlling hydrological dynamics of wetlands are functions of interacting climatic and landscape controls (Devito et al., 2005a). The western Boreal Plain, characterized by sub-humid climate, heterogeneous glacial deposits and flat terrain, has complex surface water–groundwater interactions, resulting in surface hydrological dynamics that do not necessarily reflect local topography (Smerdon et al., 2005). This landscape complexity makes it difficult to extend process-based understanding from one or two catchments to other nearby catchments or to scale hydrological findings from low-order catchments to regional catchments (e.g. surface hydrological dynamics in some catchments may be driven by topography, but in an adjacent catchment the surface hydrological dynamics maybe driven more by subsurface storage features). A complete picture of hydrological dynamics will only emerge when considering a large area that captures the full spatial heterogeneity in landscape properties and a long time to capture the multiple climatic oscillations that influence the hydrology of this region.

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Satellite data provide an opportunity for delineating surface hydrological dynamics over large geographic extents and over many years (Smith, 1997; Frazier et al., 2003). In particular, synthetic aperture radar (SAR) images have been shown to be very useful in detecting changes in soil moisture (Oldak et al., 2003; Moran et al., 2004) and inundated areas (Townsend, 2002; Bourgeau-Chavez et al., 2005). Advantages of SAR sensors (in relation to optical sensors such as Landsat TM) include all-weather, all-day and canopy penetration capabilities. The European Remote Sensing (ERS)-1 and -2 satellites have been collectively the longest running commercial systems (1992 to present) collecting fine resolution (~25 m) SAR imagery. ERS is a C-band sensor (5.6 cm wavelength) with vertical send and receive polarization and a 35-day repeat cycle (Attema, 1991). ERS has been shown to be useful in detecting soil wetness and inundation in temperate and boreal landscapes (Morrissey et al., 1994; Kasischke et al., 1995; Gineste et al., 1998; Quesney et al., 2000). ERS data are available from archives managed by various agencies (e.g. Alaska Satellite Facility (ASF)).

In this study, we monitored variations in soil saturation and inundation in a remote and relatively undisturbed part of the boreal forest. The aim was to evaluate the use of archived SAR imagery in establishing surface hydrological dynamics over a representative range of hydrological conditions. Specifically, the objectives were to: (1) map year-to-year changes in surface hydrological dynamics using ERS imagery; (2) quantify the relation between climate and hydrological dynamics; (3) quantify the change in magnitude and organization of surface hydrological patterns under a range of climatic conditions; (4) quantify the probability of surface saturation and inundation in space and time.

**STUDY AREA**

The Utikuma River drainage basin was selected as the study area (outlet located at 56°04'28.4"N, 115°08'51.9"W; Figure 1). This basin lies in the western portion of the Boreal Plain ecozone on an upland plateau, approximately 600 m a.s.l. in the larger Peace River drainage basin (Figure 1, inset). It contains Utikuma Lake, a large regional lake with a surface area of 288 km² (Mitchell and Prepas, 1990). The climate is continental with cold, long winters and short, cool summers. Based on the 1971–2000 normals for the closest year-round meteorological recording station (Slave Lake, 55°18'18", 114°46'46"W), located within 80 km of the centre of our study area, average annual temperature is 1.7°C, with average monthly temperatures ranging from −14.5°C in January to 15.6°C in July (Environment Canada, 2006). Average annual precipitation $P$ is 503 mm, 70% of which falls as rain from May to September (Environment Canada, 2006). The highest average daily precipitation occurs between early June and mid July (~3.5 mm day$^{-1}$), followed by a relatively drier period from mid July to the end of August.

![Figure 1. Map of the Utikuma River drainage basin with locations of soil moisture sampling sites marked. Inset: location of study region on the Boreal Plain ecozone](image-url)
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Figure 2. Average daily precipitation based on the 1970–2004, 35-year normal for Lesser Slave Lake, the closest year-round meteorological recording station, located within 80 km of the centre of the study area (≈2.1 mm day⁻¹) (Figure 2). Average annual potential evapotranspiration (PET) is approximately 517 mm. During the past 35 years P was less than PET (based on 1 May to 31 August totals that were available for the Utikuma River drainage basin), except once every 5 to 10 years when P was greater than PET (Figure 3). This has resulted in an overall drying trend since the mid 1970s (Figure 3).

The Utikuma River drains an area that is characterized by flat to gently undulating glaciated terrain with depositional landforms (including moraines, glacio-fluvial outwash, and glacio-lacustrine clay plains) (Paulen et al., 2004). Most flat areas are covered by peatlands, and many rivers and lakes are surrounded by swamps or marshes. Soils in poorly drained locations are Organic soils, whereas soils on better drained till and outwash deposits are dominated by Orthic and Podzolic Gray Luvisols (Wynnyk et al., 1963).

Wetland vegetation is dominated by black spruce (Picea mariana (Mill.) B.S.P.) and tamarack (Larix laricina (Du Roi) K. Koch) in peat-forming fens and bogs and by willow (Salix spp.), birch (Betula spp.), alder (Alnus spp.), sedges (Carex spp.) and grasses in non-peat-forming swamps and marshes (Mitchell and Prepas, 1990). Tree stands on the lowlands are relatively short (<4 m) with open canopies. Vegetation on moderately well-drained uplands is dominated by trembling aspen (Populus tremuloides Michx.) with occasional stands of white spruce (Picea glauca (Moench) Voss) (Mitchell and Prepas, 1990). Vegetation on well-drained uplands is dominated by pine (Pinus banksiana Lamb.). Aspen stands have closed but simple (one-layer) canopies ranging from 8 to 15 m in height. Pine stands have more open canopies ranging from 7 to 13 m in height.

METHODS

Preprocessing archived ERS imagery

ERS-1 and ERS-2 images were acquired through the ASF. All images were acquired in the descending orbit at noon local time. The images were orthorectified using a 25 m pixel resolution digital elevation model and an additional 20 to 30 ground control points collected from georeferenced geographical information system (GIS) layers of hydrography (i.e. lakes and streams), infrastructure (i.e. roads) and a recent Landsat TM image (i.e. road intersections). The root-mean-square error was <25 m in both Northing and Easting directions for all images. During orthorectification, radar images were resampled from 12.5 m to 25 m pixel resolution using a bilinear resampling algorithm and then filtered twice using a 3 × 3 gamma filter to reduce speckle (Lopes et al., 1993). We assumed that the resolution of the images (25 m × 25 m) would be sufficient to resolve the saturated and inundated features of the landscape that are often many times the size of one pixel given the relatively flat landscape.

Once orthorectified, digital numbers were converted to backscatter coefficient (with units of decibels) using standard radiometric calibration techniques (ASF, 2004). ERS-1 images were applied an offset of 0.5 dB to make them comparable to the ERS-2 images (W. Albright (ASF), personal communication). ERS-2 images were corrected for a constant rate of loss in transmitter pulse power by applying a linear correction factor (Meadows et al., 2004). The rate of loss was approximately 0.66 dB year⁻¹ from launch (1995) to the end of 2000 and then
approximately 0.82 dB year\(^{-1}\) from 2001 to February 2003 (Meadows et al., 2004). In February 2003, a gain increase of 3 dB was applied. Since then the rate of loss has been 0.66 dB year\(^{-1}\). Finally, for both ERS-1 and ERS-2 images, a systematic offset was observed between images taken from the two adjacent descending orbits, which is most likely due to differences in relative proportions of lake pixels introducing different levels of auto-normalization (Meadows et al., 1998). An offset of −0.8 dB was applied to images acquired from the orbit centred on 56°N, 116°W.

**Interpreting synthetic aperture radar imagery for hydrological applications**

SAR images capture the amount of incident microwave radiation backscattered by a surface. On landscapes where interferences are minimal (e.g. bare non-agricultural soils), the intensity of this backscattered radiation is highly dependent on the amount of water present within the top soil layer.

There are two dominant scattering mechanisms at work as the hydrological state changes from dry through to saturated and then to inundated soils (Figure 4). The first mechanism, over non-inundated areas, is controlled by the dielectric properties of the surface. Dielectric properties are related to the way a medium reacts when an electric field is applied to it (Ulaby et al., 1996). Since the dielectric constant of wet soils (~80) is much higher than the dielectric constant of dry soils (~1), the microwave radiation is very sensitive to changes in wetness (Ulaby et al., 1996). The relation between the backscatter coefficient and wetness is non-linear because the backscatter coefficient reaches signal saturation at high wetness levels (Kasischke et al., 1995). Surface roughness and vegetation may change this basic backscatter response in complex ways, sometimes by increasing the backscatter coefficient (Griffiths and Wooding, 1996) or sometimes by decreasing the backscatter coefficient (Kasischke et al., 2003). The mixed-wood boreal forest of the Utikuma River drainage basin has a relatively simple and open canopy, giving good penetration capability; therefore, we considered interferences to the backscattering signal minimal. For example, for the northern portion of the Utikuma River drainage basin, vegetation density data from the Alberta Vegetation Inventory indicated that, of the total land area, 47% has a canopy closure of 50% or less, 89% has a canopy closure of 70% or less, and only 12% has a canopy closure of greater than 100%. These data suggest that vegetation will have a relatively minor effect on the backscatter response for the majority of the watershed.

The second scattering mechanism, over inundated areas, is controlled by specular reflection from smooth water surfaces. Specular reflection implies that most of the incident energy is reflected away from the sensor and, therefore, that the backscatter coefficient will be very low from these areas. This basic backscatter response is modified where there is emergent vegetation over inundated areas. In such cases, double-bounce scattering from the water surface and plant stems substantially increases the backscatter coefficient (Townsend, 2002). The backscatter coefficient may also be enhanced over inundated areas if there is significant wind-induced wave action (Horritt et al., 2003). This can sometimes increase the backscatter signal of a lake to match that of non-inundated areas. Potential wind effects on open water areas of lakes in the Utikuma River drainage basin were reduced by excluding all open water areas of lakes as defined by a provincial hydrography layer.

**Developing a hydrological classification for ERS images**

We wanted to assess the nature of the relation between satellite-based backscatter coefficient and ground-based volumetric soil moisture (VSM) over non-inundated areas. Ideally, we would have measured VSM directly using a gravimetric method (e.g. by oven-drying soil samples) and related it to the backscatter coefficient. However, direct sampling of VSM was not possible, as we needed to cover a large geographic region in a short period of time. Therefore, we estimated VSM using a soil impedance probe and a calibration function that related gravimetrically determined VSM and soil impedance. The indirect measure of VSM may have introduced error in the estimation of VSM, but collecting soil moisture measurements with a probe had the advantage of covering the range of sites we needed.

To calibrate soil impedance (millivolts) to VSM (per cent) we collected soil samples and corresponding soil impedance measurements from representative upland and lowland sites. Twenty sites were sampled during summer of 2004, with four sites sampled twice. Soil impedance measurements were taken with a ThetaProbe (Delta-T Devices, 1998). The ThetaProbe measures the impedance of an electromagnetic wave travelling along a central rod surrounded by three outer rods, inserted 6 cm into the ground surface (Munoz-Carpena, 2002). The impedance probe was inserted into the top 6 cm of the soil, which consisted mostly of a leaf, fibric, and humic (LFH) layer in uplands and peat or moss in lowlands. For the
same soil volume (30 cm$^3$), a soil sample was taken and its moisture content determined gravimetrically. A regression model was generated to relate soil impedance and gravimetrically determined VSM (Figure 5). Since both upland (LFH) and lowland (organic) samples fell on the same regression line that related VSM to soil impedance, only one regression model was used ($r^2 = 0.74$; SE = 0.42; $p < 0.0001$, $n = 24$). To calibrate backscatter coefficient (decibels) to VSM, we extracted backscatter coefficients and coincident soil impedance measurements across 18 sites (Figure 1) during four sampling periods in the summer of 2004 (26 June, 12 July, 16 August, and 20 September). Data collection started the evening prior to and continued through to the afternoon after the satellite flyover. The sites were selected to represent the dominant cover types of the region (i.e. black spruce and tamarack forests in lowlands, aspen and pine forests in uplands). Each site was located within a larger homogeneous region (minimum area of 1 ha) with respect to vegetation cover to ensure minimal mixed pixels and to reduce noise introduced by speckle when co-registering with satellite imagery (Griffiths and Wooding, 1996). At each of the 18 sites, soil impedance readings were collected on a rectangular grid (24 nodes laid out over a 50 m $\times$ 50 m area) (Figure 1, inset). At each node of the 50 m $\times$ 50 m rectangular grid, impedance readings were taken in each of the cardinal directions, with a set of four readings in upland sites and two sets of four readings (one in a hollow and one on a hummock) in lowland sites. Soil impedance measurements were averaged for each of the 18 sites and subsequently converted to VSM using the calibration function shown in Figure 5. Corresponding satellite-based backscatter data were extracted from four ERS images by digitizing polygons on the computer screen with the aid of geographic coordinates of sampling sites and a land cover map derived from a recent Landsat TM image. An average backscatter coefficient was calculated for each polygon. Regression models were developed that related ground-based VSM to satellite-based backscatter coefficient. Prior to developing the regression models, the VSM data (which were left skewed) were transformed by taking the natural logarithm to meet the assumption of normally distributed regression residuals. Following convention, the backscatter coefficient was regressed against VSM. All four models were statistically significant and explained between 31 and 67% of the variance. When the four samples (each representing the 18 sites sampled during four different sampling periods) were combined into one global dataset, the regression model captured 45% of the variation in backscatter coefficient ($p < 0.0001$; SE = 0.90, $n = 72$) (Figure 6). Although statistically significant, the regression model explained less than 50% of the variation in radar backscatter, which we considered inadequate for characterizing VSM content. However, our targets were the mapping of inundated and saturated parts of the landscape. To map these areas, a regression model that provided a continuous measure of soil moisture was not needed; rather, a simple classification of unsaturated, saturated, and inundated classes was developed. We used the regression model (shown in Figure 6) to transform all pixels from decibels to VSM. Next, we applied classification rules to separate the derived VSM layer into three hydrologically relevant classes. For non-inundated parts of the landscape, we defined two classes: unsaturated (0–60% VSM) and saturated (>60% VSM). Our threshold in VSM for identifying saturated areas was based on the gravimetric samples collected for both mineral and organic soils. To complete the classification we defined inundated areas as anything less than 60% VSM. Our classification rules were implemented using a sigmoidal

![Figure 5](image-url)  
Figure 5. Regression model relating soil impedance measured by ThetaProbe and VSM estimated gravimetrically from soil samples. The data points represent 20 soil moisture sampling sites, 11 of which overlap with the 18 sites shown in Figure 1. The data used in the figure were collected during the July and September collection campaigns with four of the sites sampled twice, giving a total sample size of 24

![Figure 6](image-url)  
Figure 6. Regression model relating ground and satellite-based estimates of soil moisture based on four sampling periods conducted during the summer of 2004. Each data point represents one of the 18 soil moisture sampling sites shown on Figure 1. VSM was derived from soil impedance measurements using the regression shown in Figure 5.
membership function (Burrough and McDonnell, 1998), implemented in TAS 2.07 (Lindsay, 2005). The membership function is governed by the two parameters $b$ and $d$, where $b$ represents the point at which the probability of an object belonging to that class is 0.5 and $d$ represents the width of the fuzzy zone between the 0.5 and the 1.0 probability point of a particular class. The $b$ and $d$ values were based on the class threshold determined from field data and reflected the uncertainty in separating the classes. Although not shown on the graph, the fuzzy zones do intersect

We assessed the accuracy of our hydrological classification using two methods. First, we assessed the statistical separability of the classes. We grouped each soil moisture sampling site into unsaturated or saturated based on the 60% VSM threshold and we used the provincial lake layer to identify inundated areas. For each of the three classes, we extracted backscatter coefficients and tested the separability by running a Kruskal–Wallis $H$ test, which compared the mean ranking of the groups. The Kruskal–Wallis $H$ test indicated statistically significant differences among the three classes ($\chi^2 = 14737; p < 0.0001$; Figure 8). Open-water areas (without wind-induced backscatter artefacts) were used in assessing the separability of the three classes, but because of the random and, therefore, unpredictable effect of wind-induced backscatter increases in open-water areas, these areas were masked out for the rest of the hydrological pattern analyses. In contrast, shoreline and open-water wetland areas, where wind effects are reduced by nearby vegetation, were not masked out.

Second, we tested the accuracy of the classification of unsaturated versus saturated classes for 16 validation sites that were not part of the data used to develop the classification. Each of the four field collection dates was represented by these validation data. We extracted the dominant hydrological class (i.e. mode) for each site using the final classification maps and compared the classified data with the field data. The accuracy assessment indicated high overall agreement between the field-based and the satellite-based classifications (of 16 validation sites, 14 (88%) were correctly classified).
Monitoring the dynamics of surface hydrological features using archived ERS imagery

In order to capture the full range of hydrological conditions, all available descending images were acquired from the ASF archives between 1992 and 2004. One image was selected per year for the relatively dry late-summer period from year-days 200 to 240 (20 July to 30 August). This period reflects the cumulative effect of hydrological forcing on surface hydrological dynamics that result from the precipitation peak in June and July (Figure 2).

Four ERS-1 (1992–1995) and five ERS-2 (1996–2004) images were selected. We tested how representative the summer hydrological conditions during these nine ‘image’ years were of the longer 1970–2004 period of climate record. Summer hydrological conditions were approximated by summing the daily difference between precipitation $P$ and potential evapotranspiration $PET$ between 1 May and 31 August. $PET$ was computed from daily average temperature $T$ record according to Hamon (1964). $P$ and $T$ were measured at five climate stations within or near the Utikuma River basin (maximum 40 km from drainage divide). The five climate records were averaged to give a basin-wide representative climate record. $P – PET$ was calculated from this combined record. A Kolmogorov–Smirnov test showed no statistically significant difference ($z = 0.43$; $p = 0.99$) between the distribution of the $P – PET$ sample of the nine years with satellite coverage and the $P – PET$ sample of the 35-year (1970–2004) period. Closer inspection of the climate record revealed that both the second wettest (1996: 137 mm) and fourth driest (1995: –164 mm) years over the past 35 years occurred between 1992 and 2004. These results suggest that the seasonal climate totals of the nine available ERS-image years were representative of the previous 35-year (1970–2004) record.

For the analyses of surface hydrological dynamics, we combined saturated and inundated classes into a wetland class, resulting in a binary (non-wet versus wet) image. For each of the nine images, the percentage cover of wetlands ($%WET$) within the Utikuma River drainage basin was calculated (approximately 10% of the Utikuma River drainage basin did not have satellite coverage and was masked out in all subsequent analyses). We related $%WET$ to other basin-wide measures of hydro-climatic indicators, including $P – PET$ and fluctuations in water levels within Utikuma Lake. The change in lake level was computed as the difference between the lake levels measured on 1 May and 31 August.

Using the wetland maps, simple landscape metrics were calculated to capture the organization of these hydrologically important areas under different climatic conditions. We computed the total number of wetland patches for the Utikuma River drainage basin and the percentage cover of wetlands that were connected to lakes ($%WETCONN$). Wetlands had to be located not more than 25 m from the shoreline of lakes to be considered connected. This buffer allowance was implemented to factor in any locational inaccuracies in the images.

Characterizing the probability of wetland occurrence

We computed the probability of wetland occurrence using an aspatial and spatial approach. With the aspatial approach, we wanted to answer the question about what the probability of observing a certain percentage of the landscape covered by wetlands was during the month of August. For example, what was the probability of finding 5% of the landscape covered by wetlands? In order to calculate this probability, we ranked $%WET$ from lowest to highest. The lowest value of $%WET$ was assigned a probability of occurrence of 100% (i.e. nine out of nine images had at least 4.4% of the landscape covered by wetlands). The second lowest value of $%WET$ was assigned a probability of 89% (i.e. 8/9), and so on. The highest value of $%WET$ was assigned a probability of 11.1%.

With the spatial approach, we wanted to answer the question about what the probability of finding any point in the landscape belonging to the wetland class was during the month of August. The probability of wetland occurrence was computed by summing the nine binary (non-wet versus wet) image layers and dividing by the sample size. This technique was used to compute the probability of inundation in boreal forests (Clark, 2004) and the probability of belonging to a depression from digital elevation models (Lindsay and Creed, 2006).

RESULTS AND DISCUSSION

Boreal landscapes occupy large and remote areas that remain relatively untouched by human disturbance. As the human footprint changes the boreal landscape, there is a need for hydrological data to establish the range of natural variability (i.e. a reference condition) and to estimate changes to this range of natural variability caused by human activities (i.e. altered condition). A challenge is to develop databases and analyses at the scales at which these human activities are being manifested. We retrieved SAR imagery from the archives of the ASF to characterize surface hydrological dynamics over a regional catchment (that included hundreds of smaller low-order catchments) and over a longer time period (1992–2004) that captured the representative range of hydrological conditions.

Wetland dynamics

A time-series and scatter plot of $%WET$ and corresponding changes in lake levels are presented in Figure 9a and b. Both of these graphs suggest significant temporal coherence of surface hydrological patterns within the terrestrial and the aquatic systems of the Utikuma River drainage basin.

An examination of climatic controls on $%WET$ and lake level was conducted (Figure 10). On the graph, a $1:1$ line is drawn that reflects a lake level response (millimetres) that is identical in magnitude to inputs of $P – PET$ (millimetres). We observed that, when $P < PET$, an increase in $P – PET$ resulted in an equal increase...
in lake level and the data points fell parallel to the 1:1 line. However, when \( P \geq PET \), an increase in \( P - PET \) resulted in a greater increase in lake level such that the data points fell above the 1:1 line. There was a two to three times increase in lake level for each unit increase of \( P - PET \); for example, for a \( P - PET \) increase of 137 mm, the lake level increased by 630 mm in 1996. This suggested that as \( P - PET > 0 \) mm, the storage capacity within the terrestrial system started to fill, and excess water was transferred via surface pathways (i.e. streams and wetlands) and possibly subsurface pathways to the lake.

However, not all years fit the general pattern observed in Figure 10. In some years (1971, 1979, 1994) the lake
levels were more responsive than anticipated based on the $P - PET$ data. This suggests that the storage capacity was already near full due to precipitation in previous years and/or that the timing of precipitation during the course of the summer was more important than the total seasonal precipitation. The daily precipitation record during these years indicated that there was a 45 mm or greater precipitation event in all of these years during the month of June. In fact, from 23 to 26 June 1971, 115 mm of rain fell, making it the largest 4-day total over the 35-year climate record. In contrast, lake levels during 2000 were less responsive than anticipated, suggesting that the storage capacity was not reached within the terrestrial system. The small hydrological response may be explained by two exceptionally dry years (1998, 1999) that preceded 2000, which had the effect of substantially increasing the storage capacity. Scatter around the pattern observed in Figure 10 may also be attributed to the fact that PET does not represent actual evapotranspiration (AET) well in all years. Devito et al. (2005b) showed that, for a first-order catchment on the Boreal Plain, AET matched PET fairly well in wet years but was less than PET during dry years. In a regional basin where there are relatively more open water areas, AET is more likely closer to PET even in drier years.

The surface hydrological dynamics in the Utikuma River drainage basin showed the following characteristics:

1. there was a general trend showing immediate responsiveness of wetland size and water depth in lakes to climatic forcing;
2. there was evidence for a threshold in $P - PET$ ($\approx 0$ mm), below which vertical hydrological fluxes were important and above which both vertical and horizontal fluxes were important;
3. there was evidence for the importance of within-year timing of precipitation (e.g. extreme rain events during June or July) and/or among-year timing of drought and flood years (e.g. multiple years of $P < PET$) that resulted in changes to the amount of water stored on the landscape leading to time lags between climate forcing and associated hydrological response.

**Wetland connectivity**

A time-series of maps of unsaturated, saturated or inundated areas of the Utikuma River drainage basin was generated from archived SAR imagery. These maps showed a complex pattern, with saturated and inundated areas distributed along topographic divides rather than around streams and shorelines (e.g. Figure 11). This pattern suggested that surface hydrological dynamics may be regulated more by subsurface (e.g. stratigraphy) than surface (e.g. topography) controls. The importance of substrate control on the movement of water in this relatively flat region was emphasized by Devito et al. (2005a), who provided evidence for surface and subsurface water flow that did not always follow topographic drainage patterns.

There was substantial variation in wetland connectivity as hydrological conditions changed from dry to wet (Figures 11 and 12). The total number of wetland patches increased initially, peaked when %WET was approximately 15%, and then decreased gradually as %WET reached the limit of 35% (Figure 12). The connectivity of these wetland patches (%WET$_{CONN}$) increased slowly, and then more rapidly, suggesting that a threshold was reached in the water storage capacity of the contributing land areas (Figure 12). From these landscape patterns in wetlands, the following general observations can be made. When climatic inputs were small ($P \ll PET$), patterns of wetlands were disorganized and disconnected from lakes. In this dry state, there were few patches (0.08 ha) with small areas (<1 ha). When climatic inputs were balanced by outputs ($P = PET$), patterns of wetlands were still disorganized but were distributed in newly formed patches. Patch density peaked in this mesic transition state (0.13 ha$^{-1}$). When climatic inputs were large ($P \gg PET$), patterns of wetlands became organized and connected to lakes. In this wet state, wet patches were relatively few (0.10 ha$^{-1}$) but were large (>4 ha).

The trends observed in Figure 12 suggested that surface hydrological dynamics in the Utikuma River
Figure 12. Scatter plot showing relations between percentage wetland cover, number of wetland patches and percentage wetland connected to lakes.

drainage basin have two dominant states: a dry, disconnected state, and a wet, connected state. Such dual systems have been observed in other environments (Grayson et al., 1997; Cook, 2001; Western et al., 2001). Grayson et al. (1997) reported that soil moisture patterns switched between two preferred states in temperate regions of Australia. During the dry state, vertical fluxes of water dominated and the pattern of surface saturation was seemingly random, whereas during the wet state, horizontal movement of water became more dominant and the pattern of saturation became more organized (Grayson et al., 1997). Similarly, Cook (2001) reported that vertical fluxes of water dominate during dry periods, but temporary surface connections between adjacent wetlands were observed during wet periods in temperate regions of central North America.

The surface hydrological dynamics on the Utikuma River drainage basin have important implications for the movement of nutrients. Lakes may receive relatively little surface hydrological and biogeochemical inputs from their catchments during the frequent dry conditions, supporting findings from field investigations conducted in small, low-order catchments on the Boreal Plain (Ferone and Devito, 2004). That study suggested that most of the surface hydrological action takes place in riparian zones surrounding shallow lakes and wetlands. However, lakes become more connected to wetlands within their catchments and may receive larger hydrological and biogeochemical inputs during the less frequent periods of wet conditions (Devito et al., 2000; Evans et al., 2000).

We have assessed the surface hydrological dynamics in a regional drainage basin; however, there may be large sub-basin differences between climatic forcing and magnitude and organization of surface hydrological features due to a variety of landscape factors, as hypothesized by Devito et al. (2005a). For example, in groundwater discharge zones, connectivity of wetlands may be more stable over time than in areas with limited groundwater upwelling. Current research is focusing on understanding groundwater and surface water interactions in the dominant hydro-geological units of this region (e.g. Smerdon et al., 2005).

Probability of wetland occurrence

Surface hydrological dynamics captured by archived ERS imagery in the Utikuma River drainage basin were synthesized to compute the probability of wetland occurrence. The computed probabilities of wetland occurrence were based on the probability of observing wetlands after the peak in the rainy season (late July and August) between the years 1992 and 2004. However, as shown in the ‘Monitoring the dynamics of surface hydrological features using archived ERS imagery’ section, the years of this study period captured the 35-year extremes in climatic forcing; therefore, the argument can be made that the probability of wetland occurrence based on the nine study years would be valid over a longer period.

A plot of %WET against the probability of occurrence indicated the probability of observing a certain percentage of the landscape being covered with wetlands (Figure 13). For example, at any given time the probability of observing 5% of the landscape being covered with wetlands was 100%. The probability decreased as %WET increased. There was a more drastic decrease in probability until %WET reached about 12%. This inflection point reflected the transition zone in wetland reorganization when going from the dry state to the wet state (Figure 12).

A map of probability of wetland occurrence exhibited a pattern where most of the Utikuma River drainage basin had a low probability of being a wetland (Figure 14a). However, the eastern portion of the basin (dominated by clayey soils and larger peatland complexes) showed more areas with a higher probability of being a wetland (Figure 14a). Figure 14b is an example of three small lakes surrounded by an extensive peatland located on the eastern side of the Utikuma River drainage basin. The probability map showed that most of the areas surrounding the lakes had a high probability of being...
a wetland; however, the pattern was not uniform. The areas of highest probability straddled topographic divides, which illustrates that topography may not be the primary control on patterns of soil wetness (Devito et al., 2005a). More research is needed to identify the landscape controls on this observed pattern.

The synthesis of surface hydrological dynamics into probability maps provides a useful tool in identifying parts of the landscape that are most susceptible to human disturbance, such as saturated and inundated areas. The identification of such hydrologically sensitive areas for improved land management has received recent attention (Walter et al., 2000; Agnew et al., 2006). Wolniewicz et al. (2002) mapped the return period of surface saturation using radar imagery and used it to provide a...
buffer design that was adaptive to the probability of formation of hydrologically sensitive areas. Accurate delineation of hydrologically sensitive areas may also facilitate improved road placement and lead generally to ‘surprise-free’ forest operations (Tague and Band, 2001). Future work needs to focus on how best to incorporate maps of probability of wetland occurrence into forest management plans on the Boreal Plain.

CONCLUSIONS

The Utikuma River regional drainage basin, located on the Boreal Plain of northern Alberta, exhibited surface hydrological dynamics that switched between two states: a dry state characterized by a low aerial coverage of wetlands that were not well connected to lakes and a wet state characterized by much higher aerial coverage of wetlands that were well connected to lakes and other wetlands. We mapped wetlands (areas of saturation and inundation) over a 12-year period using ERS images acquired after the peak in the rainy season. The mapping was based on a simple classification where the boundaries between the classes were calibrated by ground-based soil moisture measurements. Analyses of the patterns of wetlands indicated that, during dry climatic conditions, wetland sizes were small and disconnected from each other and receiving bodies of water. As climatic conditions changed from dry to mesic, wetland numbers increased but were still disconnected from the rest of the landscape. It required very wet climatic conditions before the disconnected wetlands coalesced and connected to the lakes. During these wet conditions, the response of the lake level at Utikuma Lake was observed to be much higher than under drier conditions. Analyses of individual wetland maps and integrated probability maps may be useful in identifying not only important sites for land-to-lake horizontal hydrological linkages, but also land-to-atmosphere vertical linkages. These maps have the potential to inform future scientific studies, as well as management practices, on the Boreal Plain.

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CHARACTERIZING HYDRODYNAMICS ON BOREAL LANDSCAPES


