Observations and modeling of turbulent fluxes during melt at the shrub-tundra transition zone 1: point scale variations
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ABSTRACT
Vegetation has a significant influence on snow accumulation and energy availability for snowmelt. This is particularly true in the vicinity of the arctic treeline, characterized by the alternation of shrub-tundra and open-tundra, with the former expected to spread more and more. This work considers the time variation in turbulent fluxes over two open-tundra and shrub-tundra sites, where measurements of sensible and latent heat fluxes over the canopy are available. An improved version of the GEOtop hydrological model with a dual-layer surface scheme has been used to interpret and reproduce the measurements. The model allows us to separate the contribution of the vegetation and the surface to the turbulent fluxes measured above the canopy and, despite some issues related to the parameterization of the turbulence in the canopy, is able to reasonably reproduce the turbulent fluxes measured above the vegetation and the snowmelt acceleration observed in the shrub-tundra. The maximum energy contribution to the surface during snowmelt is found to occur for values of the leaf and stem area index around 1.0. The model proves to be a valuable platform to be applied in a distributed model to predict the spatial variability of snowmelt and surface energy balance.

Key words | Arctic regions, GEOtop, hydrological modeling, shrub-tundra, snowmelt energetics

INTRODUCTION
The transfer of water and energy between the surface and the atmosphere is greatly influenced by the vegetation cover in most environments. This is also true in many cold environments, where vegetation has a significant influence on snow accumulation and energy availability for snowmelt and soil thawing. As a consequence, vegetation heterogeneities strongly contribute to the spatial variability in snow cover and energy availability, and, therefore, the hydrology of large regions of the Arctic. This is particularly true in the vicinity of the Arctic treeline, which is characterized by dramatic changes in vegetation from northern boreal forest to shrub-tundra and open-tundra over relatively short distances (Marsh et al. 2008). As air temperatures over much of the Arctic have risen over the last few decades, there have been large changes in vegetation (Sturm et al. 2001a,b), and it is expected that the transition from tundra to shrub-tundra will occur most rapidly in the coming years as air temperatures continue to rise (Strack et al. 2007). In order to predict the effect of such changes in vegetation on hydrology and regional climate, there is a strong need to properly account for the fluxes of water and energy in these environments in hydrologic, weather and climate models.

The effect of shrub-tundra systems on snow accumulation depends on the proportion of the area covered by shrubs. Patchy shrub areas, and the edges of extensive shrub areas, have increased snow water equivalent (SWE) compared to winter snowfall as the shrubs act as a sink for blowing snow from tundra areas. Larger shrub areas typically have a snow cover that is similar in SWE as winter snowfall, due to limited transport of snow from the shrubs...
and reduced sublimation (Liston et al. 2002). However, the
effect of shrubs on snow accumulation is complicated by
the fact that they are often bent over and buried by snow
(Marsh et al. 2003). In this situation, the accumulation in
snow is often intermediate between wind-blown tundra and
shrub-tundra. In addition, shrubs significantly alter the
surface energy balance, changing the relative importance of
its components. They shade the surface, but their stems are
heated as a result of high absorption of solar radiation. As a
consequence, they supply heat to the snow surface both
directly, by long-wave radiation, and indirectly, by sensible
heat, warming the canopy air, which, in turn, provides
sensible heat to the surface. These effects cause more energy
to be available at the surface and, therefore, snowmelt is
accelerated (Pomeroy et al. 2006).

Although many studies have focused on measurement and
modeling of the snow surface energy balance components in
forested sites (e.g. Harding & Pomeroy 1996; Pomeroy & Dion
1996; Niu & Yang 2004; Marks et al. 2008; Pomeroy et al. 2008),
fewer studies have considered the surface energy balance of
shrub tundra. Shrubs present some peculiarities, related to
their short height and the fact that their characteristics change
very rapidly in time, as a result of the process of snow
emergence. Liston et al. (2002) applied a blowing snow and
energy balance model on Arctic shrub tundra, but they focused
primarily on their effect on snow accumulation, rather than on
Scheme on subarctic shrub tundra sites to consider radiative
and turbulent fluxes, accounting for the effect of canopy gaps
on solar radiation transmissivity, and using a dual-layer
surface scheme. He found that the model is not fully satisfying
in its ability to reproduce the turbulent heat fluxes at the
surface and concluded that shrub effects on the turbulent
exchange are not well understood. The problem is made more
difficult because it is not possible to measure turbulent fluxes
under shrubs, as normally done in forest (e.g. Marks et al.
2008). However, Bewley (2006) showed the importance of
accurately modelling the turbulent fluxes in order to predict
snowmelt. In addition, Marsh et al. (2010) used CLASS
(Verseghy 1991) to describe the snow surface energy balance
at the Arctic treeline, and again demonstrated the need to
properly account for the role of shrubs.

The main purpose of this paper, the first of a two-part
paper, is to consider variations in turbulent fluxes over the
two dominant vegetation types found in the vicinity of the
Arctic treeline, and to consider whether a dual-layer surface
scheme properly describes the turbulent fluxes over both
shrub tundra and open tundra during, and immediately after,
snowmelt. Due to the expected changes in vegetation in
Arctic regions in the coming decades, there is a need to better
understand the weaknesses and strengths of the dual-layer
surface approach, and to use its theoretical framework, in
combination with observations, to understand how tall
shrubs and low open tundra vegetation affect the heat and
water vapor transport to the surface. Field observations will
be used in conjunction with the hydrological model GEOtop
(Rigon et al. 2006; Endrizzi 2009), using a dual-layer surface
scheme, to perform this study. An additional goal is to verify
GEOtop in point mode so that it can be applied with high-
resolution grids in order to consider basin-wide fluxes. This
is required in order to carry out the second part of this paper
(Endrizzi et al. in preparation), which will integrate tower
flux measurements and basin-wide aircraft flux data (Mauder
et al. 2008), in conjunction with GEOtop simulations, in
order to consider the spatial variability in turbulent fluxes at a
range of scales from point to 100 km².

FIELD OBSERVATIONS AND MODELLING

Field observations

The Trail Valley Creek (TVC) research basin, located
north-east of Inuvik, Northwest Territories, Canada, at
approximately 68°45′N, 133°30′W, has observations
since the early 1990s. Full details on the study site and
instrumentation are available in Marsh et al. (2008) and
Marsh et al. (2010).

The area is dominated by gently rolling hills that are
incised by occasional steep-sided river valleys. TVC is
located in the forest–tundra transition zone, with vege-
tation consisting primarily of upland open tundra and shrub
tundra. Micro-meteorological measurements were obtained
at two primary instrumented sites in order to be represen-
tative of the main vegetative covers. These stations are TUP
(TVC Upper Plateau) and TTS (TVC Tall Shrubs). TUP is
located in the highest part of the basin, in a wide and level
area of open tundra (at 179 m a.s.l.), characterized by very
low vegetation (dwarf birch, fusicose lichens, northern Labrador tea and cottongrass). TTS is located in the largest contiguous area (approximately 2 km²) of tall shrubs (at 156 m a.s.l.), not dissimilar from the more extensive areas of shrub tundra lying further south, with vegetation consisting of green alder and diamond leaf willow, approximately 1.5 m high (Marsh et al. 2010). Figure 1 shows a map of the basin, including the location of the meteorological stations. Fetch for the TTS site is 560 m to the north and west, 330 m to the east, and 2,000 m to the south, while fetch at TUP is greater than 600 m in all directions. Due to the size of the TTS site, edge effects are small.

Measurements of classical meteorological variables (namely incoming short-wave and long-wave radiation, wind speed and direction, temperature and relative humidity) have been collected above the vegetation cover at both TUP and TTS since 2002. In addition, measurements of sensible and latent heat fluxes have been obtained through a sonic anemometer and hygrometer. However, given the high level of maintenance required by this instrumentation, measurements of turbulent fluxes have normally been performed only during the melting and summer seasons. Since the most complete dataset of measurements of turbulent fluxes is available for the 2003 snowmelt season, this study will focus on that year. However, observations and simulations for the 2005 spring seasons are considered for TTS in order to better appreciate both the inter-annual variability and the ability of GEOtop to account for such differences.

In 2003 incoming short-wave radiation was measured at TUP with an Epply B&W pyranometer (model 8–48), while incoming long-wave radiation was measured at TTS above the canopy with an Epply pyrogeometer. The radiation components were assumed to be equal at both sites, as in Marsh et al. (2010), and this is supported by the observation that measurements of incoming radiative

Figure 1: Map of Trail Valley Creek, showing the land cover and the location of the TTS and TUP meteorological stations.
fluxes were very similar in both sites in later years, including 2005, when a CNR1 Kipp&Zonen radiometer was installed in both the sites. Sensible heat fluxes were measured with a CSAT3 Three Dimensional Sonic Anemometer (Campbell Scientific), together with a KH20 Krypton Hygrometer (Campbell Scientific) for the latent heat fluxes, installed at a height of 3.2 m at TUP and 6.48 m at TTS. Due to the remote location, and since the instruments remained unattended for most of the time, only 30 min averages, variances and covariances of the 10 Hz raw data were collected. Therefore, quality control filters, like QC (Vickers & Mahrt 1997), could not be applied. However, a number of corrections have been applied on the covariances in order to remove (1) the mean vertical motion (Kaimal & Finnigan 1994); (2) the errors on the latent heat flux due to changes in the air density when sensible heat flux occurs (Webb et al. 1980); (3) the oxygen absorption at the frequencies emitted by the krypton lamp in the hygrometer and (4) the attenuation of the high frequency eddies contributing to the fluxes (Horst 1997). However, this resulted in a modest variation of the average fluxes during snowmelt (0.75 and 2.50 W/m² at TTS, and 0.12 and 0.5 W/m² at TUP, for the latent and sensible heat fluxes, respectively), though locally the corrections can be more significant. This is in accordance with other studies carried out over snow (Reba et al. 2009). After snowmelt, the corrections are more significant (10.07 and 7.86 W/m² at TTS, and 5.2 and 0.2 W/m² at TUP, for the latent and sensible heat fluxes, respectively), and a higher relative increment of the latent heat fluxes is observed as a result of a significant air density correction due to the high values of the sensible heat fluxes. Outliers, spikes and anomalous oscillations of the covariances were manually removed. In addition, it was chosen to remove the data for which the removal of the vertical motion caused an anomalous very high rotation, as a consequence of high lateral covariances.

A simple 1D footprint model (Kljun et al. 2004) was run for the maximum and average values of the friction velocity and the variance of the vertical velocity. The footprint of the TUP tower resulted in having a peak at 72 m and 51 m upwind, and its 90% dimension resulted in being 199 m and 140 m upwind, respectively, using maximum and average values. The TTS tower, as a consequence of the higher elevation of the EC sensors and a higher turbulence due to higher roughness length, resulted in having a peak footprint at 97 m and 73 m upwind, and a 90% dimension of 266 m and 200 m upwind. Therefore, the footprint results should always be included in the fetch.

**GEOtop**

**Model description**

Although there are numerous hydrological and land surface models that allow the computation of surface heat exchanges (e.g. CLASS, Verseghy 1991; VIC, Liang et al. 1994; CLM, Oleson et al. 2004), we will use the GEOtop model (Rigon et al. 2006; Endrizzi 2009) in this study. GEOtop has a number of important advantages required for meeting the objectives of this two-part study, including: (1) it is a gridded model that can be used at small grid sizes (meters to tens of meters), thus allowing the consideration of whether small scale variability is important in controlling fluxes at the tower scale (this paper) and basin scale (i.e. 100 km² discussed in the second part of this paper), (2) it calculates the full surface energy balance of both snow-covered and snow-free terrain, (3) it considers the role of vegetation on snow accumulation and melt, and has a dual-surface scheme to represent turbulent fluxes in vegetated terrain, (4) it includes a coupled blowing snow model (Pomeroy et al. 1995) for re-distributing snow fall in open terrain, (5) it uses a coupled numerical solution of the heat and water flow equations in the soil and the snow cover, thereby allowing a consideration of frost table depth, soil moisture and runoff on surface energy fluxes and (6) it has a scheme for distributing atmospheric conditions (temperature, wind, radiation) over the study area (Liston & Elder 2006). The use of such a model has the advantage that it can be used in combination with detailed observations to allow an improved understanding of the controlling processes and the disadvantage of requiring detailed initialization data. This is not a serious limitation in the current study as a large amount of data is available for the TVC research basin considered in this study. Details of GEOtop, and the field observations, used in this paper follow, while other details pertaining to the basin-wide application of GEOtop will be provided in the second part of this paper (Endrizzi et al. in preparation).
The GEOtop hydrologic model employs a finite-difference numerical method to discretize the soil and snow into as many layers as required to properly consider subsurface processes. The surface energy balance is given by the sum of shortwave radiation, longwave radiation, the turbulent fluxes of sensible and latent heats, and changes in soil or snow heat storage. Forcing data include air temperature and humidity, wind speed, incoming shortwave and longwave radiation. All of these are normally provided by measurements. However, when incoming longwave radiation measurements are not available, it is estimated using one of the numerous empirical relations existing in the literature (e.g. Brutsaert 1975) and considering a cloud correction (Sicart et al. 2004). The turbulent fluxes of sensible and latent heat are calculated according to the classical Monin–Obukhov similarity theory, with a more detailed vegetation scheme described in the next subsection.

Vegetation scheme

Previous versions of GEOtop did not consider a partitioning of the radiative or turbulent exchange energy fluxes between the surface and the canopy, but instead considered only a single homogeneous surface as a source of heat and vapor, including both the ground surface and the vegetation, considered at the same temperature (Monteith 1963). This approach often provides unrealistic results when the actual temperature and albedo of the canopy and the surface noticeably differ (Deardoff 1978), as is the case of snow-covered terrain in spring time, when the vegetation canopy and stems may warm many degrees above 0°C, while the snow surface is less than or equal to 0°C. In such cases, energy fluxes are best described by differentiating the environment near the surface from that in the canopy (McNaughton & Van den Hurk 1995). A simple method to do this is to use a two-layer approach (Choudhury & Monteith 1988), which considers heat and vapor sources both in the canopy and at the surface.

In this version of GEOtop, vegetation is treated as a homogeneously distributed mass above the surface, able to absorb and reflect shortwave radiation, to absorb, reflect and emit longwave radiation, and where air can flow and transfer heat, water vapor and momentum. A potential limitation in this version is that no vegetation gaps are considered. The vegetation density is described by the leaf and stem area index (LSAI), and its temperature \( T_v \) is calculated solving the vegetation energy balance, namely

\[
C_v \frac{dT_v}{dt} = SW_v + LW_v - H_v - LE_v
\]

where \( t \) is time, \( C_v \) the thermal capacity of vegetation, \( SW_v \) and \( LW_v \) the shortwave and longwave radiation absorbed by vegetation, and \( H_v \) and \( LE_v \) the sensible and latent heat fluxes from the vegetation to the canopy air. Equation (1) is nonlinear, since \( LW_v \), \( H_v \) and \( LE_v \), in turn, depend on \( T_v \), and is solved with a globally convergent Newton–Raphson method (Tomita 2009).

The transfer of shortwave radiation within the canopy is calculated with the two-streams approximation of Dickinson (1985) and Sellers (1985). Longwave radiation is calculated assuming that the canopy emits radiation equally towards the soil and towards the atmosphere, and considering a vegetation emissivity that increases with vegetation density.

The main assumption of the dual-layer surface scheme is that the air within the canopy has a negligible capacity to store heat and water vapor. Therefore, the heat and vapor fluxes from the stems and canopy (respectively named \( H_v \) and \( E_v \), named canopy fluxes) and from the surface (respectively \( H_s \) and \( E_s \), named undercanopy fluxes) are respectively balanced by the heat and vapor fluxes above the vegetation (\( H \) and \( E \), named total fluxes). In particular

\[
H_v = \rho c_p \frac{T_v - T_{ca}}{r_v} \quad H_s = \rho c_p \frac{T_s - T_{ca}}{r_{uc}} \quad H = \rho c_p \frac{T_{ca} - T_a}{r_{ah}}
\]

\[
E_v = \rho q_v - q_{ca} \quad E_s = \rho q_s - q_{ca} \quad E = \rho q_{ca} - q_{a}
\]

where \( T_{ca} \) (or \( q_{ca} \)) is the temperature (or specific humidity) of the air within the canopy, \( T_s \) (or \( q_s \)) is the temperature (or specific humidity) at the surface, be it snow-covered or snow-free, \( T_a \) (or \( q_a \)) is the temperature (or specific humidity) of the air above the canopy, \( q_v \) the specific humidity at the canopy surface, assumed to be saturated, \( r_{ah} \) (or \( r_{ae} \)) is the above-canopy aerodynamic resistance for heat (or water vapor) transfer, \( r_v \) (or \( r_s \)) is
the aerodynamic resistance of turbulent transfer of heat (or water vapor) between the canopy and the canopy air (named canopy resistances), \( r_{uc} \) is the aerodynamic resistance of turbulent transfer of both heat and water vapor between the soil or snow surface and the canopy air (named undercanopy resistances, after Zeng et al. (2005)), \( \rho \) is the air density and \( c_p \) the specific heat at constant pressure. Since \( H = H_v + H_s \) and \( E = E_v + E_s \) for the above-mentioned assumption, \( T_{ca} \) and \( q_{ca} \) are dependent on the other variables and are weighted averages of the air, canopy and surface temperatures, and specific humidities.

An accurate determination of the resistances is essential in this approach. The above-canopy aerodynamic resistances are determined with the classical Monin–Obukhov similarity theory, assuming typical values of the ratios of roughness length for momentum to vegetation height, and vegetation zero-displacement height to vegetation height (Lindroth 1993). In the dual-layer surface scheme the roughness lengths for heat and water vapor transport are considered equal to the roughness length for momentum. The differences normally considered between the roughness lengths are related to the existence of a sublayer very close to the surface, where momentum and scalar transport take place in different ways, since molecular transfer and pressure fluctuation are important for momentum, while only molecular transfer controls heat and water vapor transport (Cahill et al. 1997). This effect can also be described assuming the roughness lengths for scalar equal to the roughness length for momentum, and adding an additional resistance (Zeng & Dickinson 1998), which can be seen as the limit of the above-mentioned undercanopy resistance \( r_{uc} \) when the canopy disappears and reduces to a bare soil formulation (Zeng et al. 2005).

On the other hand, much more uncertainty exists in the determination of the under-canopy aerodynamic resistance, due to the poor understanding of the under-canopy turbulence (McNaughton & Van den Hurk 1995). This is particularly true in the sparse canopy case, which is, in addition, the most sensitive to the parameterizations of the resistances (Zeng et al. 2005). The aerodynamic resistance \( (r) \) to momentum, sensible heat flux and evaporation between the heights \( z_1 \) and \( z_2 \) is defined as (Choudhry & Monteith 1988):

\[
r = \int_{z_1}^{z_2} \frac{dz}{K(\varepsilon)}
\]

where \( \varepsilon \) is the vertical coordinate and \( K \) the eddy diffusivity. The general approach assumes an exponential decay of the eddy diffusivity within the canopy, namely

\[
K(\varepsilon) = K(H_c) \cdot \exp \left[ -n \left( 1 - \frac{\varepsilon}{H_c} \right) \right]
\]

where \( H_c \) is the canopy height and \( n \) a decay coefficient. The value of the eddy diffusivity at the canopy height level is determined by the Monin–Obukhov theory

\[
K(H_c) = k u^* (H_c - d) (\Phi(H_c))^{0.5}
\]

where \( k \) is the Von Karman constant, \( u^* \) is the friction velocity, \( d \) is the vegetation zero-displacement height and \( \Phi \) is the stability function above the canopy resulting from the application of the Monin–Obukhov theory. Integrating (4) between the roughness length of the surface (\( z_{0v} \)) and the height at which the sink of momentum for the effect of the canopy is considered to occur (\( d + z_{0v} \)), one obtains

\[
r_{uc} = \frac{H_c}{nK(H_c)} \left\{ \exp \left[ n \left( 1 - \frac{z_{0v}}{H_c} \right) \right] - \exp \left[ n \left( 1 - \frac{d + z_{0v}}{H_c} \right) \right] \right\}
\]

Commonly, \( n \) ranges from approximately 2.5–3.0 in crops (Goudriaan 1977) and 2.5–4.0 in forest (Brutsaert 1982). However, some authors assume that for sparse vegetation a lower decay coefficient may apply. For example, Zeng et al. (2005) implicitly assume

\[
n = 0.7 - \text{LSAI}
\]

which may provide values remarkably lower than the range 2.5–4.

In an attempt to use a more physically based formulation, Niu & Yang (2004) propose

\[
n = \left( \frac{c_d H_c}{L_m} \right)^{0.5} \left( \Phi_c \right)^{0.5}
\]
where \( c_d \) is the drag coefficient, \( l_m \) is the canopy mean mixing length (i.e. the free space between leaves and stem) and \( \Phi_c \) is a stability correction factor within the canopy, where stability conditions different from above the canopy may occur within the canopy according to the temperature gradients between the surface and the canopy air. However, this formulation is quite difficult to implement for many types of vegetation (including open tundra and shrub tundra), because of the uncertainty in the estimation of \( c_d \) and \( l_m \). Given this consideration, in this work the following formulation, similar to, but simpler than, (9) has been used:

\[
n = n_0(\Phi_c)^{0.5}
\]

(10)

where \( n_0 \) is a parameter depending on the vegetation type.

The determination of the aerodynamic canopy resistance is less critical, since it covers a more moderate range in values (Huntingford et al. 1995). However, the canopy fluxes depend linearly on LSAI, so a more accurate determination of this parameter is actually more important than finding a more sophisticated formulation of the resistances. The canopy latent heat fluxes are given by the component due to wet vegetation and that due to transpiration. On the latter the stomatal resistance has to be considered.

The formulation of the resistances proposed here relies on the theory of local diffusion (K theory), namely on the assumption that temperature and water vapor concentrations only depend on local fluxes and local turbulence, and are not affected by fluxes occurring at other levels in the canopy, so that the eddy diffusivity has a local value and can be integrated as in (4) (McNaughton & Van den Hurk 1995). Actually, it has been shown that in the canopy the K theory is not fully valid, since the sources and sinks are, in fact, distributed, and the largest eddies may be responsible for non-local and even counter-gradient fluxes (Shaw & Pereira 1982). The present approach assumes that the vegetation is dense enough so that the effective source and sinks of heat and water vapor can be considered to occur at the one characteristic height, namely \( d + z_{0,v} \). Choudhury & Monteith (1988) suggest that this can be reasonable for LSAI greater than 0.8. For lower values of LSAI, Shuttleworth & Wallace (1985) suggest the use of a combination theory providing a transition between the bare soil and the closed canopy limits.

**Fractional vegetation area**

When dealing with a sparse canopy, some models introduce the concept of fractional vegetation area (e.g. the Community Land Model; Zeng et al. 2002) and consider the bare soil formulation of the surface energy fluxes in a fraction of the grid cell and the formulation given by the vegetation dual layer scheme in the complementary fraction. The surface energy balance is then a weighted average of the surface energy balance of the vegetated and unvegetated portions. This approach allows overcoming some of the problems related to the determination of the under-canopy resistance in the presence of a very sparse canopy, since it is possible to consider a denser canopy only in a part of the grid cell.

In this work we are dealing with shrub tundra vegetation, which has the peculiar characteristic of being very flexible, and it can be flexed into a nearly horizontal position, and then partially or completely buried by snow during winter (Marsh et al. 2003). This occurs even if the shrubs are much higher than the snow depth. Then, as the snow melts, the partially buried shrubs gradually emerge from the snow: the shrubs typically jump over a period of a few hours to the vertical position, but with different timing among each other, as soon as the weight of the snow cover becomes smaller than a specific threshold.

The shrub emergence process has been accounted for in this version of GETopt by using the concept of fractional vegetation area. Two threshold snow depth values have been defined: a value \( D_{11} \) above which all the shrubs are considered buried and a value \( D_{12} \) below which all the shrubs are considered unburied. When the snow depth \( D_s \) is smaller than \( D_{11} \), but greater than \( D_{12} \), the shrubs are considered to be partially emerged from the snow cover, and a fractional vegetated area \( f_v \) is defined as

\[
f_v = 1 - \left(\frac{D_{12} - D_s}{D_{12} - D_{11}}\right)^q \leq 1
\]

(11)

where \( q \) is an exponent greater than 1 (set at an attempt value of 2 in this work), which takes into account that the probability of shrub emergence occurs more rapidly as snow becomes shallower (Pomeroy et al. 2006). The differences in the dynamics of the snow burial process during the accumulation time (which becomes more intense as snow
become deeper) and the snow emergence process during snowmelt are accounted with different values of $q$.

**GEOtop initiation**

Table 1 reports the values of the parameters used for vegetation in this study. The values were taken from similar studies of shrub tundra (e.g. Sturm et al. 2001a,b; Bewley 2006; Marsh et al. 2010). A limited calibration was performed only on the parameters $n_0$, $D_{t1}$ and $D_{t2}$. However, a sensitivity analysis of the most significant parameters will be carried out to better understand their role in the partitioning of the energy fluxes between vegetation and surface (LSAI and $n_0$). Vegetation was considered non-transpiring during the melt period and for a brief period after snow is removed, and this was accounted for by assigning a very high value to the stomatal resistance.

Simulations at both TTS and TUP have been carried out from 25 April 2003 to 14 June 2003, and only at TTS from 25 April 2005 to 30 May 2005, covering the whole snowmelt period. In addition, several days after snow removal were considered in order to examine both snow-covered and snow-free conditions. The snow cover has been initialized with values provided by snow survey performed close to the sites. The initial values are reported in Table 2.

### RESULTS

The occurrence of periods when air temperatures were above or below 0°C significantly affects the fluxes of sensible heat and, to a lesser degree, latent heat. As a result, the following discussion will focus on the sequence of warm and cold spells in both 2003 and 2005 (Figures 2(a), 3(a) and 4(a)). Warm and cold spells are defined as periods when temperature rises above 0°C during daytime or always remains below 0°C, respectively. The 2003 cold spells occurred from 11–17 May and from 22–25 May, while the warm spells were from 18–21 May and after 26 May. During the 2005 melt period, the warm periods occurred from 12–15 May and after 17 May, and the cold periods before 12 May and from 16–17 May.

### Snow cover removal and shrub emergence

Although the SWE was higher at TTS than TUP at the start of melt (Table 2), the model results show snow removal occurred on 29 May 2003 at TUP and on 28 May 2003 at TTS (Figure 5). These dates are earlier than observed, but the removal of the TTS snowpack prior to that at TUP was as observed as shown by the fact that the observed snow covered area was 0.5 at TUP and only 0.125 at TTS on 30 May 2003. While the snowcover was melting, the corresponding evolution of the canopy fraction and the resultant shrub emergence is reported in Figure 6. The earlier removal of the snow cover at TTS, compared to TUP, in 2003 must have occurred due to this emergence of the shrubs and their effect on enhancing snowmelt. The details of this effect will be explored in the following sections of this paper. In the 2005 simulation, the snow removal occurred on 21 May at TTS, as there were fewer cold periods during the melt period in that year than in 2003.

The discrepancies between the observed and modeled disappearance of the snow cover in 2003 are at least in part due to the fact that GEOtop is applied here in point mode.
Figure 2 | Temperature of the air (measured) and surface and vegetation (modeled) (a), observed and modeled sensible (b) and latent heat fluxes (c) at TUP during snowmelt in 2003. The continuous vertical line shows the snow removal date.

Figure 3 | Temperature of the air (measured) and surface and vegetation (modeled) (a), observed and modeled sensible (b) and latent heat fluxes (c) at TTS during snowmelt in 2003. The continuous vertical line shows the snow removal date.
using an average snow water equivalent and, as a result, snow disappears instantaneously. In reality the snow cover becomes patchy at the small scale due to variations in the end of winter snow cover at a small scale. As a result, it is difficult to directly compare the dates of snow removal between observed and modeled, when GEOtop is run in point mode. The discrepancies noted here will be further addressed in the second part of this paper when the model is run in distributed mode.

**Sensible heat fluxes at TUP and TTS**

**During snowmelt**

At the open tundra site (TUP), the sequence of warm and cold spells results in positive sensible heat fluxes (i.e. directed upwards) in the central part of the day during the cold spells, and in negative sensible fluxes (i.e. directed downwards) during night and during the warm spells (Figure 2(b)). This occurs because the midday snow surface
is often warmer than the air during the cold spells (Figure 2(a)), due to surface heating by shortwave radiation, even if the surface temperature does not normally reach 0°C. In contrast, during warm spells, the sensible flux is negative as the air temperature rises above 0°C and the snow surface cannot rise above the freezing temperature. The modeled sensible heat fluxes at TUP agree quite well with the observations during the warm spells. On the other hand, in the cold spells the sensible heat fluxes are often overestimated (i.e. from 10 May to 15 May 2003, and 24 and 25 May 2003) when the modeled peak values are up to twice that of the observed fluxes. In addition, the model describes night minima slightly negative, while the observations are in general around 0. However, the temporal patterns of the fluxes are normally well reproduced. Reasons for these discrepancies will be considered later in the paper.

At the shrub tundra site (TTS), the observed sensible fluxes (Figure 3(b)) are positive, and the alternation of cold and warm spells (Figure 3(a)) is, in general, reflected in episodes of high and low fluxes measured above canopy, respectively. In general the observed and modeled fluxes are very similar (Figure 3(b)), even if the model overestimates on the same days (e.g. 11 and 26 May 2003). Compared to the open tundra site, sensible fluxes are higher, and on warm days they are positive instead of negative, but still smaller in magnitude than the fluxes during cold days. This occurs because the shrubs gradually emerge, warm up and, in turn, warm the canopy air close to the surface and increase the magnitude of the fluxes detected above the canopy. The largest difference in sensible fluxes between TTS and TUP occurs close to the end of the melting season, when the shrub emergence process is almost completed. The gradual increase in sensible flux at TTS compared to that at TUP from observations demonstrates the effect of the emerging shrubs. The ability of the model to reproduce the observations suggests that GEOtop is able to reproduce the effect of the shrubs’ gradual exposure on the sensible heat fluxes, and that the dual layer scheme provides a reasonable estimate of the surface contribution (i.e. under-canopy fluxes) and the vegetation contribution (i.e. canopy fluxes), and therefore the fluxes above vegetation (i.e. total fluxes). On the cold days, both vegetation and snow surface temperatures are higher than the air temperature above the canopy, and, consequently, they both contribute with positive components, whereas at TUP only the surface component exists (Figures 2(b) and 3(b)). On the other hand, on warm days, vegetation temperature rises above the air temperature, while the snow surface is bound by the freezing point. Therefore, total sensible fluxes above the vegetation are a combination between positive canopy fluxes and negative under-canopy fluxes.

Since only the undercanopy component contributes to the surface energy balance at TTS and, therefore, to snow melting, it is worthwhile to compare the under-canopy fluxes at TTS with the total sensible fluxes at TUP, which directly contribute to the snow energy balance since no, or very little, vegetation emerges during most of the snowmelt period. As the TTS surface directly exchanges heat with the canopy air, which is generally warmer than the air above the canopy as a result of the warming effect of the shrubs, with respect to TUP, more sensible heat is conveyed to the TTS surface in warm spells, and less heat is removed from the TTS surface in cold spells. Therefore, the overall effect of shrubs is to reduce sensible heat losses and increase sensible heat gains, thus increasing the total sensible heat flux to the surface and enhancing melt compared to the tundra site. This effect is clearly enhanced as the shrubs are continually exposed during melt, and is also consistent with previous studies (Pomeroy et al. 2006).

Table 3 shows a summary of the modeled and observed turbulent fluxes both at TUP and TTS in 2003, reporting...
measured and simulated averages and root mean squares (RMS). It also shows some quantitative measures of the goodness of fit, in particular the mean of bias (MB) equal to the mean of observations minus the mean of simulations, the root mean square error (RMSE), and the Nash–Sutcliffe efficiency, given by 1-RMSE/RMSobs^2, where RMSobs is the RMS of observations. In order to perform a consistent comparison, for the instants for which the measurements were removed or not available, the corresponding simulated values were not considered in the calculation.

### After snowmelt

After snowmelt, a marked increase of the sensible heat fluxes is observed (Figures 7(b) and 8(b)) at both TTS and TUP. If the daily peak values are taken as a reference, the observed fluxes are much higher at the shrub tundra site than at the open tundra site: while at TUP maximum daily measured values normally range between 200–300 W/m², peak values between 350–500 W/m² are common at TTS. The modeled fluxes agree quite well with the observations at TUP, but at TTS they are underestimated by approximately 100 W/m² at the peak. However, the general temporal patterns of the fluxes are reproduced. Table 3 shows that MB over the period from snow removal date to 14 June 2003 is 4.31 W/m² (5.4% of the observed mean) at TUP and 216.01 W/m² (12.9% of the observed mean) at TTS. NSE is 0.83 at TUP and 0.88 at TTS, which means that the fit is very good. However, NSE must be considered with caution, since high values of NSE can be more easily obtained when RMS of the observation is high, as in this case (McCuen et al. 2006).

The model is able to predict larger fluxes at the shrub site, even if not as much as observed. Actually, the low value of LSAI at both sites results in little difference between the modeled vegetation and surface temperatures (Figures 7(a) and 8(a)), which are normally 5–15°C higher than the measured air temperature in the central part of the day for both the sites. The modeled surface temperature is slightly higher at TUP and, hence, slightly higher temperature gradients occur at this site, which, however, result in lower sensible heat fluxes. The difference in modeled sensible

<table>
<thead>
<tr>
<th>Summary of measured and simulated averages and root mean squares (RMS) of sensible (H) and latent (LE) heat fluxes from 9 May 2003 to the snow removal date, and from snow removal date to 14 June 2003, at TUP and TTS. The error indices include the mean of bias (MB) equal to the mean of observations minus the mean of simulations, the root mean square error (RMSE), and the Nash–Sutcliffe efficiency, given by 1-RMSE/RMSobs^2, where RMSobs is the RMS of observations. In order to perform a consistent comparison, for the instants for which the measurements were removed or not available, the corresponding simulated values were not considered in the calculation.</th>
<th>Measured</th>
<th>Simulated</th>
<th>Error Indices</th>
</tr>
</thead>
<tbody>
<tr>
<td>H at TUP snowmelt</td>
<td>0.55</td>
<td>1.89</td>
<td>23.38</td>
</tr>
<tr>
<td>H at TTS snowmelt</td>
<td>17.08</td>
<td>17.90</td>
<td>37.64</td>
</tr>
<tr>
<td>LE at TUP snowmelt</td>
<td>7.06</td>
<td>6.82</td>
<td>14.94</td>
</tr>
<tr>
<td>LE at TTS snowmelt</td>
<td>12.04</td>
<td>15.69</td>
<td>22.67</td>
</tr>
<tr>
<td>H at TUP snow free</td>
<td>75.71</td>
<td>80.02</td>
<td>85.85</td>
</tr>
<tr>
<td>H at TTS snow free</td>
<td>124.31</td>
<td>108.3</td>
<td>122.74</td>
</tr>
<tr>
<td>LE at TUP snow free</td>
<td>37.88</td>
<td>28.64</td>
<td>28.42</td>
</tr>
<tr>
<td>LE at TTS snow free</td>
<td>24.56</td>
<td>27.48</td>
<td>22.93</td>
</tr>
</tbody>
</table>
Figure 7 | Temperature of the air (measured), and surface and vegetation (modeled) (a), observed and modeled sensible (b) and latent heat fluxes (c) at TUP after snowmelt in 2003. The continuous vertical line shows the snow removal date.

Figure 8 | Temperature of the air (measured), and surface and vegetation (modeled) (a), observed and modeled sensible (b) and latent heat fluxes (c) at TTS after snowmelt in 2003. The continuous vertical line shows the snow removal date.
heat fluxes at the open tundra and shrub tundra sites is then to be attributed to the larger vegetation height at TTS, which results in a more efficient turbulent exchange. However, apparently the shrubs enhance turbulent transfer at a higher rate than modeled.

Latent heat flux at TUP and TTS

During snowmelt

Periods of high positive latent heat fluxes (i.e. directed upwards or evaporation) can be recognized at both the sites during the cold spells (Figures 2(c) and 3(c)), and periods of low positive or slightly negative fluxes (i.e. directed downwards or condensation) during the warm spells. The fluxes are, in general, higher at TTS, and this is more evident as the shrub emergence process proceeds. The model is able to reproduce the temporal patterns of the observations, especially on the cold days, and the results are consistent with the fact that fluxes at TUP are lower than at TTS.

At the end of snowmelt (after 27 May 2003) at both the sites, the model simulates negative fluxes (condensation), while the observations still show positive values (evaporation). This may be due to a fractional snow-covered area in the footprint area of the eddy correlation sensor, with probably positive fluxes from the snow-free part, which may compensate the condensation on the snowpack. This issue will be further addressed in the next paper where the model is run in distributed mode. On 26 May 2005 the model provides very high fluxes (above 100 W/m² at the peak) at TTS, which are actually not observed. This is due to high snow evaporation from the canopy, as a result of a snowfall occurring on the day before, which deposited some snow on the shrubs. The intercepted snow has probably been overestimated. In addition, on the cold days the model overestimates the fluxes and predicts fluxes approximately twice as large as the observations from 12–15 May 2003 for both sites, but the agreement becomes better at the end of snowmelt (from 21–25 May 2003). Table 3 shows that MB is lower at TTS (2.92 W/m²) than at TUP (−9.24 W/m²), where it is relatively high and negative as a result of the underestimation of the peaks. NSE indicates a good fit, but shows a lower agreement than for the sensible heat fluxes. The model reproduces the relative difference in magnitude of the latent heat fluxes at the two sites because the specific humidity at the ground surface at TTS is higher at TUP, as a result of higher surface temperatures (Figures 7(a) and 8(a)) and, therefore, higher humidity gradients at TUP are established. The more efficient turbulence at TTS does not significantly affect the water vapor transport: since vegetation does not transpire yet, the only source of water vapor is the ground surface, which, given the high hydraulic conductivity of the peat soil (Quinton et al. 2000), dries out quickly, and, therefore, even if the under-canopy resistance becomes quite low, the soil resistance to evaporation and, consequently, the overall resistance remain high.

After snowmelt

The observed latent heat fluxes after snow removal (Figures 7(c) and 8(c)) are prevalently positive (i.e. directed upwards), and are much lower in magnitude than the correspondent sensible heat fluxes, displaying maximum values of the order of 100 W/m². The measured latent heat fluxes in the central part of the day at TTS are up to two-thirds higher than at TUP. The model is able to reproduce the observation of higher fluxes at TUP, even if with a lesser degree than the measurements. However, the latent fluxes are underestimated in some days, especially at TUP, in particular on the days when the air temperature is lower (after 10 June 2003). Table 3 shows that MB is lower at TTS (2.92 W/m²) than at TUP (−9.24 W/m²), where it is relatively high and negative as a result of the underestimation of the peaks. NSE indicates a good fit, but shows a lower agreement than for the sensible heat fluxes.

The model reproduces the relative difference in magnitude of the latent heat fluxes at the two sites because the specific humidity at the ground surface at TTS is higher at TUP, as a result of higher surface temperatures (Figures 7(a) and 8(a)) and, therefore, higher humidity gradients at TUP are established. The more efficient turbulence at TTS does not significantly affect the water vapor transport: since vegetation does not transpire yet, the only source of water vapor is the ground surface, which, given the high hydraulic conductivity of the peat soil (Quinton et al. 2000), dries out quickly, and, therefore, even if the under-canopy resistance becomes quite low, the soil resistance to evaporation and, consequently, the overall resistance remain high.

SENSITIVITY ANALYSIS

Since there is a moderate degree of uncertainty in the value of some of the vegetation parameters used in GEOtop
(LSAI), or a lack of understanding of the physical processes (decay coefficient of the eddy diffusivity in the canopy), a sensitivity analysis has been carried out to assess their influence on the representation of the turbulent fluxes and the surface energy balance. The effect of the inclusion of the stability correction within the canopy (Equation (10)) is also studied.

Sensitivity tests include:

(i) LSAI: 4 values of LSAI: 0.5, 1.0 (base simulations), 1.5 and 2.5 at the TTS location for both study years. Values of LSAI equal to 0.3, 0.1 and absence of vegetation are also considered to better understand the effect of the shrub vegetation on the surface energy balance.

(ii) Decay coefficient of the eddy diffusivity in the canopy ($n_0$, as defined in Equation (10)): this parameter appears in the definition of the under-canopy aerodynamic resistance and, therefore, affects the turbulent transport of heat and water vapor to the surface. The sensitivity analysis is very important, since the definition of this parameter is empirical, and many formulations exist in the literature (Goudriaan 1977; Niu & Yang 2004; Zeng et al. 2005). The following values of the decay coefficient have been here considered: 0.7 (as suggested by Zeng et al. (2005)), 1.5, 2.5 (base simulation) and 3.5, so covering a large part of the variability for this parameter.

(iii) Under-canopy stability functions: values of the decay coefficient of the eddy diffusivity profile in the canopy equal to 0.7 and 2.5 with and without the stability correction defined in (10).

Since different values of vegetation parameters may cause a delay or an advance of the time of snow removal, it would be difficult to examine the influence of a single parameter on the turbulent fluxes. In order to overcome this problem, the snow cover has been forced to disappear at the same instant as in the simulations described in the previous paragraphs, referred to as base simulations.

**LSAI**

As LSAI increases, the net radiation at the surface is characterized by a change in its wavelength spectrum. The under-canopy shortwave component decreases, because of greater solar radiation absorption by the canopy, while the undercanopy longwave component increases, because of larger emission by the canopy in the infrared bands (Pomeroy et al. 2006). The present simulations show that, for the study area and for the set of parameters chosen, the cumulative net radiation during snowmelt increases with LSAI until LSAI is 1.0, and then decreases (Figure 9).

In this sensitivity analysis, we force the snow to be removed at the same time in each simulation in order to facilitate the comparison. Therefore, differences in LSAI result in different values of the total energy available for snowmelt, which can actually be seen as if the initial snow depth was tuned so that snow always disappears on the same day. The energy available for snowmelt (Figure 9) shows the same dependence on LSAI as net radiation, which also demonstrates that in shrub tundra vegetation more energy is available for snowmelt than in the case of no vegetation, as in the case of open tundra. The sensible heat fluxes during snowmelt are, on average, negative (i.e. they act as a net energy source for the surface) and increase with LSAI, but they are not able to counterbalance the effect of net radiation when LSAI is larger than 1.0. On the other hand, they accentuate the dependence of the total surface energy on LSAI if this parameter is less than 1.0. The latent heat fluxes are positive (i.e. they act as a net energy sink), and are practically unaffected by LSAI changes, except for very low values of LSAI.

**Decay coefficient of the eddy diffusivity in the canopy**

The effect of decreasing the value of this parameter is to increase the under-canopy turbulence, which results in a large increase in magnitude of the undercanopy sensible heat fluxes (Figure 10(a, e)) and latent heat fluxes (Figure 10(d, h)), regardless of the directions.

This effect is especially evident when events of high sensible fluxes towards the snow surface occur at the end of snowmelt (22, 26, 27 and 28 May 2003, and from 18–21 May 2005): the magnitude of the sensible fluxes is more than doubled from $n_0 = 2.5$ to $n_0 = 0.7$, with the consequence of strongly increasing the snowmelt rate. A value as low as −300 W/m² can occur at the peak (27 May 2003), even if the under-canopy stability correction defined in (10)
has been included. The latent heat fluxes also exhibit this sharp increase in magnitude during these days but, since they are positive, they tend to offset the contribution of sensible heat fluxes.

The canopy sensible fluxes (Figure 10(b, f)) also increase in magnitude, but remain positive, as $n_0$ decreases, because the higher values of the under-canopy sensible fluxes directed to the snow surface cause a moderate cooling of the canopy air, which increases the temperature gradients at the canopy. However, the increase in the canopy fluxes is, in general, lower than the decrease in the undercanopy fluxes. As a result, in these events, the total sensible fluxes (Figure 10(c, g)) tend to assume high negative values in the daytime (26 and 27 May 2003) or at night (from 18–21 May 2005) which, if compared to the observations, seem to be unrealistic, as well as the corresponding increase in the latent heat fluxes. This demonstrates that the degree of coupling between the surface and the canopy air is quite low during snowmelt, and this is represented by relatively high values of $n_0$. This is consistent with other studies (e.g. Niu & Yang 2004; Bewley 2006).

The total energy available for snowmelt (Figure 11) varies with this parameter following the variations of the turbulent fluxes, as net radiation remains unaffected. Since the decrease in magnitude of the sensible heat fluxes, acting as energy gains, is much larger than the increase in magnitude of the latent heat fluxes, acting as energy losses, it results in a strong decrease with $n_0$ of the total energy, showing a great sensitivity to this parameter.

**Inclusion of the under-canopy stability functions**

The stability correction has an effect in the events of high sensible heat transfer towards to the snow cover at the end of snowmelt (22, 26, and 27 May 2003, and from 18–23 May 2005), because the vegetation and the canopy are warmer than the air above the canopy, while the snow surface is colder than the air above canopy, being constrained at 0°C during melting. Therefore, while the
boundary layer above the canopy is unstable, the air levels close to the surface are stably stratified, which tends to suppress turbulence and decrease the coupling between the surface and the canopy air. The canopy stability correction is intended to account for this process, but here it does not seem to have an appreciable effect, since it contributes to reduce the sensible heat fluxes, as energy gain, from 38.54 MJ (without correction) to 33.88 MJ (with correction) if \( n_0 = 0.7 \), and from 18.12 MJ (without correction) to 16.62 MJ (with correction) if \( n_0 = 2.5 \). The reason can be that in TVC wind is always quite strong (normally above 2 m/s), and this prevents the stability correction from having a great effect.

**DISCUSSION**

A dual-layer scheme proves to be able to reproduce reasonably well the temporal course of the turbulent fluxes above the canopy at both the open tundra and shrub tundra sites, during snowmelt and shortly after snow removal. In particular, it is able to describe how the shrubs, once emerged from the snow cover and warmed due to their low albedo, enhance snowmelt both providing sensible heat to the snow surface and increasing net radiation at the surface with respect to a non-vegetated configuration. However, the agreement is not fully satisfied in the following conditions:

- At the open tundra site during snowmelt, the sensible heat fluxes are overestimated during the cold days, namely when air is colder than 0°C: this means that the model tends to predict higher temperatures of the snow surface or tends to overestimate the turbulence transfer, despite the low value at which the roughness length of the snow surface has been set (0.1 mm). An analogous overestimation is found for the latent heat fluxes.
- At the shrub tundra site after snowmelt, sensible heat fluxes are normally underestimated in their peak values. This suggests that turbulent transfer is actually more
efficient than predicted by the model. Even if higher values of LSAI and vegetation height as well as lower values of the decay coefficient of the eddy diffusivity profile are used, the resulting sensible fluxes are lower than the measurements. Therefore, the underestimation may be due to intrinsic weaknesses of the dual-layer scheme and the parameterization of the under-canopy resistances. Analogous discussion applies to the latent heat fluxes.

Results also show that during snowmelt the under-canopy turbulent heat fluxes are strongly sensitive to the decay coefficient of the eddy diffusivity profile in the canopy, especially when the sensible heat fluxes are directed downwards. This strongly affects the timing of snowmelt, and has also been noted by other authors (Lee & Mahrt 2004; Niu & Yang 2004; Bewley 2006). However, a more elaborate description of turbulence in the canopy requires additional data (Niu & Yang 2004), both in terms of a more detailed representation of the canopy structure and of meteorological data within the canopy, which are rarely available in remote locations and for distributed applications. Lee & Mahrt (2004) proposed a new, more complex, formulation of the eddy diffusivity, alternative to the exponential decay, and show that, for low bush vegetation, it is able to improve the representation of the sensible heat flux in the case of snow-covered terrain.

Figure 11 | Histograms reporting the cumulative energy fluxes during snowmelt for different values of the decay coefficient of the eddy diffusivity profile in the canopy at TTS in 2003 (a) and 2005 (b). Symbols as in Figure 9.
However, it still requires more data than the classical formulation that has been used in this work.

However, snowmelt is reasonably well described with high values of the decay coefficient. In particular, Bewley (2006) obtained good results of snowmelt rates and sensible heat fluxes over subarctic shrub tundra with $n_0$ equal to 3.75, but without considering the stability correction in the canopy. Niu & Yang (2004) used $n_0$ equal to 3 in a pine forest, but introduced the stability correction, as in (9). In TVC the inclusion of the stability correction only slightly affects the results, probably because the wind speed is always rather high. Therefore, the turbulent fluxes are more sensitive to $n_0$. However, in this work, with a value of $n_0$ of 2.5, the model is able to reasonably predict the snowmelt rates and the turbulent fluxes at both the sites. Therefore, the model constitutes a valuable platform that can be applied in distributed mode to predict the spatial variability of snowmelt and the surface energy balance, which strongly influence the hydrology of cold regions, affecting water availability and soil thawing dynamics. In particular, the model can be applied to perform the second part of this work, namely the comparison of maps of turbulent fluxes provided by the model with corresponding maps derived by eddy correlation measurements performed from aircraft (Mauder et al. 2008).

CONCLUSION

In this work a new version of the GEOtop hydrologic model, including a dual-layer surface scheme, was applied in point mode at an open tundra and shrub tundra site to check its capability to reproduce the turbulent fluxes of sensible and latent heat measured above the canopy, and to consider the main processes controlling fluxes at both sites. In the shrub tundra, vegetation, initially buried by the snow cover, gradually emerges and affects the turbulent transfer of heat to the snow cover. According to the dual-layer scheme, the sensible heat flux measured above the canopy (total fluxes) can be considered due partly to the canopy (canopy fluxes) and partly to the surface (under-canopy fluxes).

During snowmelt, when air temperature is above freezing for a significant part of the day, the total fluxes at the shrub tundra site are a composition between positive (i.e. upwards) canopy fluxes, since the vegetation is warmer than the air, and negative (i.e. downwards) under-canopy fluxes, since the snow surface is at the freezing temperature, while at the open tundra only the downwards component is present. At the shrub tundra, normally, the upwards component prevails and the total sensible fluxes are positive which, on the other hand, are negative at the open tundra site. This is confirmed by the observations. Actually, the undercanopy flux at the shrub tundra site is the component that provides energy to the snow and is larger in magnitude than the sensible heat flux at the open tundra site and, therefore, it contributes to accelerate snowmelt. Analogously, it results that, in cold days during snowmelt, when air temperature is below freezing and the surface temperature is higher than the air temperature because of the incoming solar radiation, the shrubs protect the surface from losing large amounts of sensible heat. This is evidence of how shrubs modify the surface energy balance with respect to open tundra, in addition to their effect on net radiation.

However, the determination of the under-canopy sensible heat flux depends on the parameterization of the eddy diffusivity within the canopy, which in this approach was considered to decay exponentially from the values above canopy. At the considered sites the results are very sensitive to the values assigned to the decay coefficient, although a correction accounting for stability conditions in the canopy was considered. However, good results are, in general, obtained using high values of the decay coefficient. A more elaborate description of the eddy diffusivity in the canopy would be useful, but would require more meteorological data in the canopy and more information on the canopy structure.

The turbulent fluxes have been examined also for a short time after snowmelt, and it was found that the shrub tundra site presents higher sensible heat fluxes than the open tundra site, because of higher turbulence as a result of a greater vegetation height, while latent heat fluxes are higher at the open tundra.

Both during and after snowmelt, and despite the issue of the parameterization of the canopy turbulence, the model proves to reproduce the fluxes reasonably well. This shows that the model can reliably be applied at the basin scale
to reproduce the spatial variability of the surface energy fluxes at a range of scales.

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