TEMPERATURE DRIVEN VAPOR FLUXES IN SOILS CAUSE A NET RECHARGE

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RESUMEN. Los gradientes de temperatura pueden impulsar la difusión de vapor en el suelo, ya que controlan la presión de vapor. Se ha estudiado la difusión de vapor en el suelo en dos climas diferentes: un clima semiárido en El Cabril (Córdoba, España) y un clima subártico en la cuenca alta del río Tuul (Mongolia). Para El Cabril se estudiaron flujos difusivos mediante temperaturas medidas y un modelo analítico. Para el Alto Tuul se desarrolló un modelo de balance de agua y energía que simula procesos como la fusión-congelación y difusión de vapor. Los resultados muestran, que el vapor difunde hacia abajo durante el verano y hacia arriba durante el invierno, mientras que los flujos promediados difunden hacia abajo. La cantidad total es pequeña para El Cabril, pero significativa para el Alto Tuul. Estos últimos valores altos se deben a las altas oscilaciones de temperatura e Mongolia y la congelación-descongelación.

ABSTRACT. Temperature gradients can drive vapor diffusion by controlling vapor pressure in the soil. We studied vapor diffusion for soils in two different climates: A semiarid climate at El Cabril (Córdoba, Spain) and a subarctic climate in the Upper Tuul River basin (Mongolia). For El Cabril vapor diffusive fluxes were studied by means of the measured temperatures and an analytical model. For the second site (Upper Tuul) a physically based soil water and energy balance model was developed accounting for relevant processes such as melting-freezing of water and vapor diffusion in the soil. Results of both sites show that vapor diffuses downwards during summer and upwards during winter, while yearly averaged fluxes diffuse downwards. The overall amount is small for El Cabril, but significant for the Upper Tuul. The latter large values can be explained by the large temperature oscillations of the Mongolian climate and the freezing/thawing of subsoil layer.

1. INTRODUCTION

Water transfer mechanisms in the soil are essential for understanding a broad range of hydrologically relevant phenomena, ranging from water uptake by plants to aquifer recharge or flood generation. These processes can be quantified using models, which can be divided into distributed and lumped models. In distributed models, the partial differential equations governing multiphase non-isothermal flow are solved with finely discretized grids. In lumped models, balances are performed over integrated portions of the domain. The separation is relevant, because vapor diffusion, which is the focus of our work, has been studied using distributed models (Ross, 1984; Scanlon and Milly, 1984, Gran et al., 2011). The main conclusion from these works is that vapor diffusion becomes significant when there are high temperature gradients, which cause large vapor pressure gradients. This has led to the belief that soil temperature and water vapor diffusion in unsaturated soils are only relevant in arid and semiarid systems, where large surface temperatures may cause downwards vapor fluxes. We conjecture that large temperature and vapor pressure gradients may also occur under subarctic climate conditions, where (very) cold conditions may remain in the soil, while (moderately) high temperatures can be reached in the surface leading to significant downwards water and energy fluxes. In short, vapor diffusion may be a relevant water transfer mechanism for extreme climates. In spite of this, vapor diffusion is ignored in lumped schemes, which are widely used to assess water resources, probably reflecting that hydrology and soil science have been largely developed in temperate climates.

This work is motivated by the need to assess the importance of vapor diffusion as a water transfer mechanism. Specifically, it is clear that changes and oscillations in atmospheric temperature and radiation cause fluctuating temperature gradients. The question is whether vapor diffusion driven by these oscillations is relevant and leads to a net downward or upward water and energy flux. We analyze this question for two cases, a semiarid and a subartic climate, providing quantitative information about the relevance of vapor diffusion and its effect on recharge. Moreover, we make use of lumped schemes, that take into account vapor diffusion, until now ignored by such schemes.

2. THEORY AND PHENOMENON

Vapor diffusion is controlled by Fick's Law, which is written as:

$$J_v = -\frac{M}{R(T+273.15)} D \nabla p_v \quad (1)$$

Where D is the diffusion coefficient (m2 s^{-1}), T is temperature (ºC), M is the molecular weight of water (0.018 kg/mol), R is the gas constant (8.31 J mol^{-1} K^{-1}), and p_v is vapor pressure (Pa). The molecular diffusion coefficient of water vapor in air depends on several factors (notably water content and tortuosity) and it has been a subject of discussion because observed vapor fluxes tend to be much larger than predicted with Equation (1) using reasonable values of tortuosity (Cass et al., 1984; Gran et al., 2011). Since this is not

the purpose of our work, we assume tortuosity to be equal to 1, so as to compensate possible increases of vapor flux due to diffusion enhancement. Therefore, we will assume $D = \theta_g D_0$, where θ_g is the volumetric content of air and D_0 is the diffusion of water vapor in air (2.4 10^{-5} m^2 s^{-1}).

The main assumption behind our work is that relative humidity in the soil is approximately equal to 100%, which requires suctions of 1000's of meters that can only be achieved through heating. It is safe to assume that $p_v = p_{vsat}$, where p_{vsat} is the saturation vapor pressure, which is only a function of temperature (it also depends on salinity, through osmotic effects, and on suction, through the psychrometric Law, but we will neglect these dependencies). Several approximations are available for p_{vsat}. We will use the one by Murray (1967) [p_{vsat}(T) = 611 exp(17.27T/(237.2+T))]. This allows us to rewrite Fick's law by expanding the gradient as:

$$J_v = -\frac{M\theta_g D_0}{R(T+273.15)}\frac{dp_{vsat}}{dT}\nabla T \quad (2)$$

We have estimated this flux using two approximations. First, using the analytical solution to heat conduction, assuming that latent heat flux can be neglected. Second, we perform energy and water balances in the root zone to analyze the role of latent heat fluxes in subarctic soils. In both cases, the total heat flux in the soil is given by the sum of conductive and advective (essentially latent) heat fluxes:

$$J_{energ} = J_{cond} + J_{adv} = -\lambda\nabla T + (L_0 + C_v T)J_v \quad (3)$$

where C_v is the soil thermal capacity (1.93 KJ kg^{-1} ºC^{-1}), λ its thermal conductivity (typically between 0.5 and 2 Wm^{-1}ºC^{-1}) and L_0 is the latent heat of vaporization (2.5·10^6 J kg^{-1}).

3. SEMIARID CLIMATE SITE: EL CABRIL

El Cabril site, located in southern Spain, contains a heavily instrumented pilot cover. We have used soil temperature and water contents, as well as meteorological data from this site to analyze vapor fluxes. Details of this work are provided by Gran (2015).

We have used directly Equation (2), assuming that temperature can be approximated as:

$$T(t,z) = T_{my} + f_d(t,d) + f_y(t_d,d) \quad (4)$$

where z is depth below the surface, t is time, t_d is Julian day (functions of t_d are treated as constant during the day), f_i is the daily (i=d) or yearly (i=y) temperature fluctuation:

$$\begin{aligned} f_i(t,z) &= A_i e^{-\alpha_i z}\sin(\beta_i) \\ \beta_i &= \omega_i(t - t_{0i}) - \alpha_i z \end{aligned} \quad (5)$$

Where, for $i = d,y$, t_{0i} is the time with mean temperature, ω_i is the frequency ($\omega_i = 2\pi/P_i$, where P_i = 365.25 days for $i = y$ and P_i = 1 day for $i = d$), A_i is the amplitude of temperature fluctuations at the surface, and α_i is the decay (with depth) constant, inverse of damping depths (L_i), given by $\alpha_i = 1/L_i = \sqrt{(\omega_i C_v/2\lambda)}$. Daily amplitude, assumed constant during each day, is larger in summer than in winter and was assumed to also vary sinusoidally during

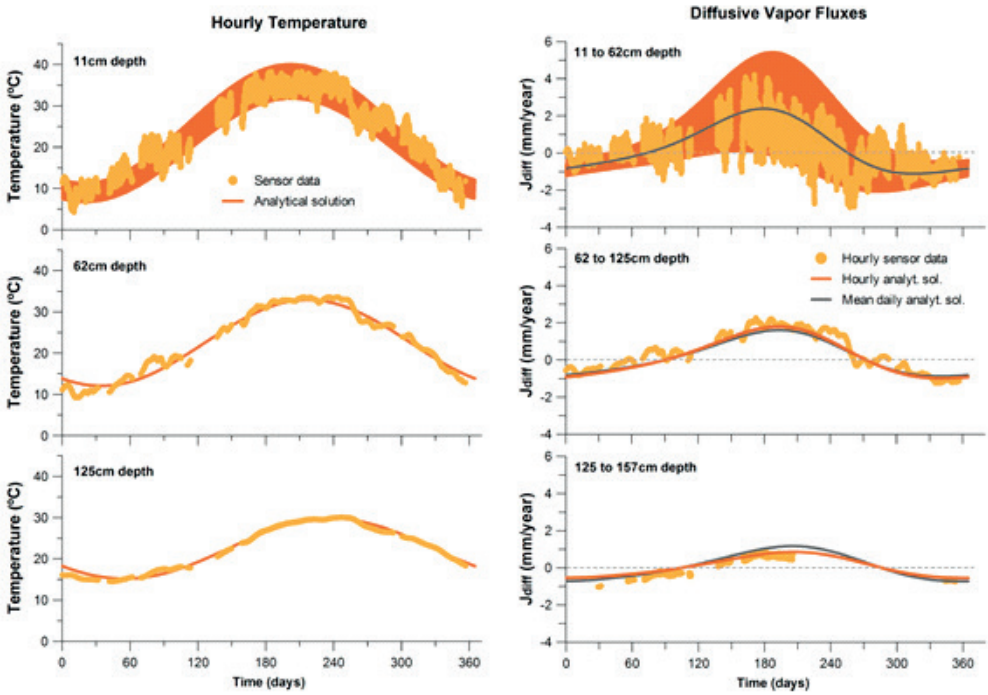

Fig. 1. Hourly temperatures (left) and vapor flux estimates (right) at several depths measured (yellow dots) and computed (lines) at El Cabril during 2009.

the year (i.e., $A_d(t_d) = A_{md} + \Delta A_d \sin(\omega_y(t_d - t_{oA}))$).

As shown in Figure 1, Equation (4) yields a good approximation of temperature fluctuations at several depths. Moreover, relative humidity measurements confirm that, except for the sensors closest to the soil surface, relative humidity remains close to 100% throughout the year. Therefore, we assume it is reasonable to use Equation (4) to estimate vapor fluxes, which yields the vapor fluxes of Figure 1. Several issues deserve discussion in this figure. First, the analytical solution (Equation 4) yields a good approximation of both mean fluxes and their fluctuations. Second, even though temperature as a sinusoidal function, vapor diffusion is tilted, with a slow increase during late winter and spring and a relatively fast drop during the fall. This drop is more abrupt in the observed data than in the analytical solution and typically occurs after a heavy rainfall. The evaporation of the infiltrated precipitation cools down the shallow soil depths, thus reversing temperature (and vapor pressure) gradients. This can be observed in mid-September (Julian day 255) in Figure 1. Third, and most important for the purpose of this work, downward vapor fluxes (positive values) are larger than the upward fluxes. This reflects the non-linear dependence of saturated vapor pressure on temperature, which causes vapor pressure to increase faster with high temperatures (summer time) than with lower ones (winter time). As a result, downward vapor pressure gradients (soil hot at the surface and cool at depth) during the summer are much larger than the upward fluxes during

the winter, thus causing a net downward vapor flux.

Table 1. Mean, maximum and minimum vapor fluxes (mm/yr) computed for two depth intervals at El Cabril during years 2009 and 2010.

Depth	11-62 cm		62-125 cm	
Year	2009	2010	2009	2010
Maximum	−2.94	−3.11	−1.2	−1.1
Mean	0.15	0.42	0.44	0.33
Minimum	4.2	5.2	2.17	2.3

Mean, maximum and minimum diffusive vapor fluxes computed from measured temperatures by means of the analytical solution (actual expressions given by Gran, 2015) are shown in Table 1. The results show that:

1. The mean yearly vapor flux is always downwards, while daily fluxes may be upwards (in winter) or downwards (in summer).
1. Mean fluxes are sensitive to mean temperature and very sensitive to temperature fluctuations (the flux increases with A_i^2).
2. The fluxes decay exponentially with depth, with a damping depth equal to half that of temperature. This implies a low penetration of daily fluctuations, but a significant one for seasonal fluctuations of vapor flux, because they can reach well below root depth.

4. SUBARTIC CLIMATE SITE: UPPER TUUL

The second case refers to the Upper Tuul River Basin around Ulaanbaatar. The region is mountainous with predominantly grassland in the south face of mountains and flat areas and predominantly forest (Larix and Pinus) in the north face of the mountains. The average daily maximum and minimum air temperature are 5.06°C and −11.5°C, respectively. Average air temperature is −3.2°C for the study period. Annual precipitation averaged is 334 mm/year, with 80% falling between June and September. In short, the region is rather cold and dry, so that it contains discontinuous and sparsely insular permafrost (Gravis et al., 1972; Sharkhuu, 2003; Jambaljav et al., 2008).

Dandar et al. (2017) developed a lumped model to perform water and energy balances over two layers: a surface layer, which represents the top 16 cm, accommodates the roots of typical grass in the basin and dampens daily temperature fluctuations, and a subsoil layer, which accommodates the "active" layer that freezes and thaws seasonally. The model accounts for the conventional water balance terms (precipitation, whether as rain or snow, evapotranspiration, including both ice deposition and sublimation and infiltration into the subsoil). Vapor diffusion was also included to assess the conjecture that it may represent a significant water transport mechanism. To this end, we discretize Equation (1) as

$$J_v = \frac{M\theta_g D_0}{R(T+273.15)} \frac{(p_{v,sf} - p_{v,ss})}{L_{sf}} \quad (6)$$

where subscript *sf* stands for surface layer, whereas *ss* stands for subsoil layer. Note that we use L_{sf} as length between the two layers rather than $((L_{sf}+L_{ss})/2$ because temperature gradients, which control vapor

pressure, are expected to be largest near the soil surface.

The energy balance considers solar radiation, latent and sensible heat fluxes, heat conduction between the two layers and energy released due to phase changes. The model also takes into account the slope and orientation of the surface. All energy fluxes can be written as a function of meteorological data and two state variables: mass of water (kg m^{-2}) and energy (J m^{-2}).

Details of the two balances are given by Dandar et al. (2017). The model was tested using meteorological data of the Terelj station (elevation 1540 m), 40 km east of Ulaanbaatar. The model was run from 2000 through 2004, using parameters from the literature and assuming that the surface is horizontal and covered by grass.

Vapor diffusion is significant. While its rate is small, it occurs throughout the late spring and summer, after the subsoil has started to thaw. Overall, it is about half of infiltration. However, Dandar et al. (2017) showed that the yearly averaged vapor diffusion is rather constant for the various years and displays little sensitivity to model parameters.Results are summarized in Table 2 and Figure 2. As usual, evapotranspiration is the main water sink, but

Infiltration occurs only after heavy rainfall events. Both infiltration (similar pattern as recharge) and downward vapor diffusion (positive in figure 2b) transform almost directly into recharge because, in the absence of deep rooted plants, the subsoil is always close to field capacity. Contrary to vapor diffusion, recharge from rainfall infiltration can vary a lot from year to year, due to the irregular occurrence of heavy rainfall events. However, a significant amount of recharge occurs throughout late spring and early summer driven by vapor diffusion into the subsoil. Note that upward vapor diffusion (negative in figure 2b) occurs during fall and early winter, but its magnitude is smaller than downward diffusion because temperatures and thus vapor pressures are also small.

Table 2. Water and energy balances averaged during 2000-2004. Precipitation is 334 kg m^{-2} $year^{-1}$.

Water balance	kg m^{-2} $year^{-1}$	Energy balance	MJ m^{-2} $year^{-1}$
Evapotransp.	284.9	Net radiation	2064.4
Infiltration	30.9	Latent heat	717.1
Surface runoff	0.1	Sensible heat	1345.6
Vapor diffusion	18.1	Vapor convection	45.9
Recharge	48.8	Soil conduction	–45.8

The energy balance terms follow the expected patterns. Latent and sensible heats in figure 2d are positives (upwards), except for heat conduction (soil heat flux) and vapor convection. Conductive heat flux is usually considered seasonal, with yearly averages close to zero. Downward heat fluxes in summer are usually balanced by upward fluxes in winter. However, even though the subsoil remains frozen for long (more than 7 months, compared to less than six the surface layer), there is a net flux upwards, to compensate the latent heat convection associated to vapor diffusion, which flows downwards. Therefore, it is not surprising that all factors that reduce the soil heat flux cause an increase in vapor convection, and vice versa. Moreover, it can be observed that the downward diffusion concentrates in the spring, when the subsoil melts (temperatures of 0°C).

An increase of the subsoil length (L_{ss}) leads to temperatures in the subsoil that oscillate less due to the increased heat storage capacity. This leads to larger tem-

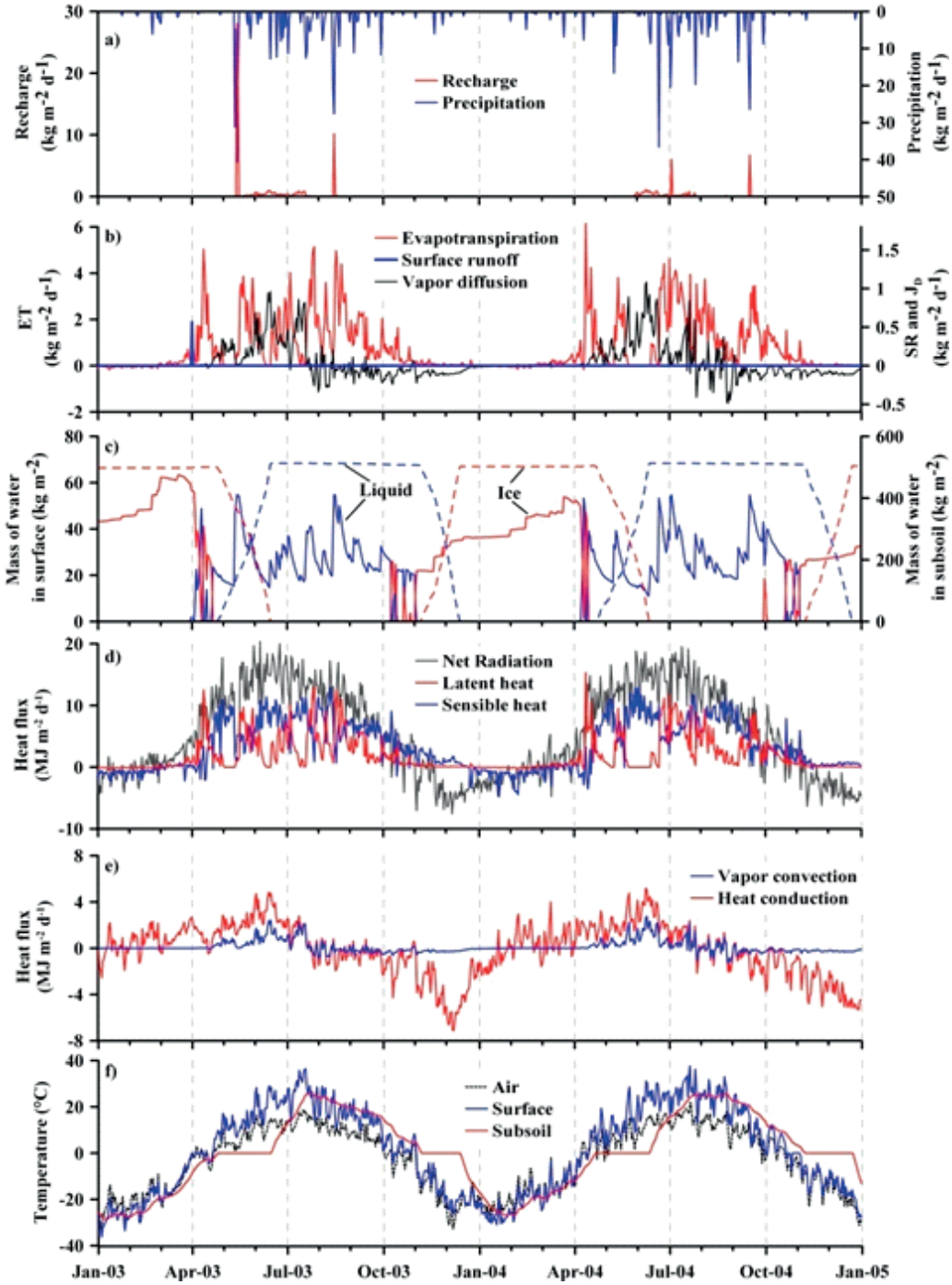

Fig. 2. Daily evolution of water fluxes, water content at the surface (solid) and subsoil (dashed), heat fluxes, and temperatures during 2003 and 2004.

perature differences between surface and subsoil, which according to our model (equation 6, Figure 2f) leads to larger vapor diffusion. As more water is transported downwards, evapotranspiration decreases and recharge increases.

5. CONCLUSIONS

The main conclusion from this work is that the mean vapor flux is downwards in both climates. This has been shown by analytical and model calculations for both sites. The flux is quantitatively small in both of them, but it is significantly larger for the Upper Tuul Basin in Mongolia, where the low magnitude of rainfall and infiltration causes vapor diffusion to be a significant source of recharge.

The nature of vapor diffusion fluxes is markedly different in both sites. At El Cabril, vapor diffusion follows loosely the seasonal fluctuations of temperature and is largest during mid-Summer. While this might look paradoxical, it is consistent with the findings of Gran et al. (2007), who observed that the condensation of water vapor diffusing downwards causes a decrease in water salinity. At the Tuul Basin, the net vapor diffusion flux is more related to the freezing/thawing of the subsoil and is also higher due to the larger temperature oscillations. Although still small, vapor diffusion may be relatively important during dry periods. In fact, during most time of the year recharge is linked to vapor diffusion.

Vapor diffusion is also important from an energy balance point of view. Heat conduction into/from the soil is often neglected because it fluctuates both daily and seasonally, so that its mean value is close to zero. As it turns out, there is a net upwards conductive flux to compensate the net downwards flux of latent heat. This is especially relevant at the Tuul Basin, because this flux controls the spring thawing of the active permafrost layer.

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