



Self-potential monitoring of a thermal pulse advecting through a preferential flow path



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SUMMARY

There is a need to develop new non-intrusive geophysical methods to detect preferential flow paths in heterogeneous porous media. A laboratory experiment is performed to non-invasively localize a preferential flow pathway in a sandbox using a heat pulse monitored by time-lapse self-potential measurements. Our goal is to investigate the amplitude of the intrinsic thermoelectric self-potential anomalies and the ability of this method to track preferential flow paths. A negative self-potential anomaly (–10 to –15 mV with respect to the background signals) is observed at the surface of the tank after hot water is injected in the upstream reservoir during steady state flow between the upstream and downstream reservoirs of the sandbox. Repeating the same experiment with the same volume of water injected upstream, but at the same temperature as the background pore water, produces a negligible self-potential anomaly. The negative self-potential anomaly is possibly associated with an intrinsic thermoelectric effect, with the temperature dependence of the streaming potential coupling coefficient, or with an apparent thermoelectric effect associated with the temperature dependence of the electrodes themselves. We model the experiment in 3D using a finite element code. Our results show that time-lapse self-potential signals can be used to track the position of traveling heat flow pulses in saturated porous materials, and therefore to find preferential flow pathways, especially in a very permeable environment and in real time. The numerical model and the data allows quantifying the intrinsic thermoelectric coupling coefficient, which is on the order of –0.3 to –1.8 mV per degree Celsius. The temperature dependence of the streaming potential during the experiment is negligible with respect to the intrinsic thermoelectric coupling. However, the temperature dependence of the potential of the electrodes needs to be accounted for and is far from being negligible if the electrodes experience temperature changes.

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1. Introduction

The self-potential method is a passive geophysical method that is remotely sensitive to any thermodynamic force affecting the motion of electrical charge carriers in a porous medium (e.g., Revil and Linde, 2006; Revil, 2007). These thermodynamic forces include the pressure head (streaming potential, e.g. Abaza and Clyde, 1969), the gradient of the concentration (diffusion potential, Mainault et al., 2005, 2006; Martínez-Pagán et al., 2010; Straface and De Biase, 2013), and the temperature gradient (thermoelectric effect, Leinov et al., 2010; Revil et al., 2013). Any of these thermodynamic forces can be the source of a source current density (Revil, 2007), which generates in turn, in the electrically conductive ground, an electrical field that can be remotely monitored.

A number of papers have been published in using self-potential signals (measured without any specific triggering effects) to identify preferential flow paths in dams and embankments (Gex, 1980; Corwin, 1985; AlSaigh et al., 1994; Sheffer and Howie, 2001; Song et al., 2005; Rozycki et al., 2006; Bolève et al., 2007, 2009, 2012). However, this purely passive mapping approach is not always characterized by a good signal-to-noise ratio. We have recently developed a better approach, called the SMART test, to illuminate preferential flow channels in dams and embankments. The SMART test consists in injecting a pulse of salty water upstream and in monitoring the advection of the salt plume using time-lapse self-potential measurements (Ikard et al., 2012). This “active” method is more precise than the conventional approach and can also be used to invert the permeability field when combined with time-lapse resistivity measurements and in situ sampling of the pore salinity in downstream wells (Jardani et al., 2013). This approach is based on the fact that the injection of a salt pulse, in a steady-state flow field, generates an electrical current density

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perturbation that can be followed intrusively (Maineult et al., 2005, 2006) or non-intrusively (Martínez-Pagán et al., 2010; Bolève et al., 2011) with the self-potential method. However, it is often forbidden, for legal reasons and water quality concerns, to inject salty water into an upstream reservoir.

The idea we follow in this paper is to extend the SMART approach by replacing the salt tracer test by a heat pulse test (T-SMART), the temperature can be eventually be lower than the background temperature. We consider the T-SMART approach more eco-friendly. In this case, the resulting self-potential signals are possibly thermoelectric in nature. In presence of a temperature field, we may also need to consider the effect of temperature on the streaming potential associated with the flow of the ground water (e.g., Somasundaran and Kulkarni, 1973; Ishido and Mizutani, 1981; Revil et al., 1999a,b). Little is known, however, about the fundamental physical principles governing the thermoelectric effect in porous media in presence of flow. Recent laboratory works on the thermoelectric effect include the work of Leinov et al. (2010) and a theoretical model can be found for instance in Revil (1999). Recently, Revil et al. (2013) have successfully used the thermoelectric effect to localize the burning front of a coal seam fire at a depth of 15 m in Colorado. While there are other geophysical methods that can identify heat pulses, like electrical resistivity tomography (e.g., Hermans et al., 2012), only the self-potential method can be used in real time to follow the heat pulse in a strongly advection-dominated system.

In this paper, we develop the concept of the T-SMART test and we apply this new test to the localization of a preferential flow channel using the injection of a heat tracer test and monitoring non-intrusively the migration of this heat pulse using the self-potential method. In Section 2, we present the theoretical background for modeling intrinsic thermoelectric potentials and their origin during heat advection and conduction through porous media. In Section 3, we describe a laboratory experiment performed to localize a preferential flow channel in a sandbox. In Section 4, we perform a numerical experiment aimed at understand the laboratory results and to provide a value to the intrinsic thermoelectric coupling coefficient.

2. Theoretical background

2.1. Ground water flow

We first present the flow equation in a partially saturated or fully saturated porous material. We solve below the generalized Richards equation with the van Genuchten parametrization for the relative permeability and the capillary pressure in an isotropic unconfined aquifer. Hysteresis will be neglected. The governing equation for the flow of the water phase is given by (Richards, 1931),

$$[C_e + s_e S] \frac{\partial \psi}{\partial t} + \nabla \cdot \mathbf{u} = 0 \quad (1)$$

where $\mathbf{u} = -K \nabla(\psi + z)$ denotes the Darcy velocity (in m s^{-1}), z and ψ denote the elevation and pressure heads, respectively, $C_e = \partial \theta / \partial \psi$ denotes the specific moisture capacity (in m^{-1}), θ is the water content (dimensionless), s_e is the effective saturation, S is the (poro-elastic) storage coefficient at saturation (m^{-1}), and t is time (in s). The effective saturation is related to the relative saturation of the water phase by $s_e = (s_w - s_w^r) / (1 - s_w^r)$ where $\theta = s_w \phi$ and ϕ (dimensionless) represents the total connected porosity of the material. The hydraulic conductivity K (in m s^{-1}) is related to the relative permeability k_r (dimensionless) and to the hydraulic conductivity at saturation K_s (m s^{-1}) by $K = k_r K_s$. Using the van Genuchten–Mualem model (van Genuchten, 1980; Mualem, 1986), the effective saturation, the relative permeability, the specific moisture capacity, and the water content are defined by,

$$s_e = \begin{cases} \frac{1}{[1 + |\alpha \psi|^n]^m}, & \psi < 0 \\ 1, & \psi \geq 0 \end{cases} \quad (2)$$

$$k_r = \begin{cases} S_e^L \left[1 - \left(1 - S_e^{\frac{1}{m}} \right)^2 \right]^2, & \psi < 0 \\ 1, & \psi \geq 0 \end{cases} \quad (3)$$

$$C_e = \begin{cases} \frac{\alpha m}{1-m} (\phi - \theta_r) S_e^{\frac{1}{m}} \left(1 - S_e^{\frac{1}{m}} \right)^m, & \psi < 0 \\ 0, & \psi \geq 0 \end{cases} \quad (4)$$

$$\theta = \begin{cases} \theta_r + s_e (\phi - \theta_r), & \psi < 0 \\ \phi, & \psi \geq 0 \end{cases} \quad (5)$$

respectively, where θ_r is the residual water content ($\theta_r = s_w^r \phi$), and α , n , $m \approx 1 - 1/n$, and L are parameters that characterize the porous material (van Genuchten, 1980; Mualem, 1986). All these parameters will be independently determined for the experiment reported in Section 3. Thermo-osmosis and electro-osmosis correspond to the influence of the thermal gradient and electrical field on the Darcy velocity, respectively. We consider these effects as negligible (see Sill, 1983 for a discussion of these effects).

2.2. Heat flow

We assume below that local thermal equilibrium between the fluid and solid phases is reached at any time. Indeed the characteristic time to reach thermal equilibrium for a silica grain of diameter d immersed in a background of uniform temperature is $\tau = d^2 / (4\alpha)$ where $\alpha = \lambda_s / (\rho_s C_p)$ denotes the thermal diffusivity of silica ($\alpha = 1.4 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$), λ_s corresponds to the thermal conductivity of silica, ρ_s its mass density, and C_p its heat capacity at constant pressure. This yields in turn a characteristic time of 0.3 s for $d = 1.4 \times 10^{-3} \text{ m}$ (grain diameter for the coarse channel, see below). As this characteristic time is much smaller than the characteristic time associated with the transport of the heat pulse in the tank ($>10 \text{ min}$), the assumption of local thermal equilibrium is checked.

We consider the flow of heat in partially saturated porous material described by Eq. (6) obtained by combining Fourier's law (the constitutive equation for the heat flux) with the continuity equation for heat,

$$\nabla \cdot (-\lambda \nabla T + \rho_w C_w T \mathbf{u}) + \rho C \frac{\partial T}{\partial t} = Q \quad (6)$$

where T is the average temperature of the porous medium (in K), λ (in $\text{W m}^{-1} \text{ K}^{-1}$) is the thermal conductivity of the porous material, ρ_w and ρ (kg m^{-3}) denote the mass density of the pore water and the bulk mass density of the porous material (with the pore fluids), respectively, C_w and C (in $\text{J kg}^{-1} \text{ K}^{-1}$) are the heat capacity of the pore water per unit mass and the bulk heat capacity of the porous material (with the pore fluids) per unit mass, respectively, and Q denotes the heat source (in W m^{-3} , positive for a source and negative for a sink). For unsaturated porous media, we determine the bulk heat capacity per unit volume and the bulk thermal conductivity by (e.g., Luo et al., 1994),

$$\rho C = (1 - \phi) \rho_s C_s + \phi s_w \rho_w C_w + \phi (1 - s_w) \rho_a C_a \quad (7)$$

and

$$\lambda = \lambda_s^{1-\phi} \lambda_w^{\phi s_w} \lambda_s^{\phi(1-s_w)} \quad (8)$$

respectively, and where C_s , ρ_s , and λ_s denotes the volumetric heat capacity, density, and thermal conductivity of phase ξ (solid s , pore water, w , and air, a).

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