Inverse Modeling for Variably Saturated Water Flow Coupled with Heat Transport in Field Soil

IZUMI Tomoki*, FUJIHARA Masayuki*, TAKEUCHI Junichiro**
and KAWACHI Toshihiko**

* Faculty of Agriculture, Ehime University, 3-5-7 Tarumi, Matsuyama-city, Ehime 790-8566, JAPAN.
** Graduate School of Agriculture, Kyoto University, Oiwake-cho, Kitashirakawa, Sakyō-ku, Kyoto-city, Kyoto 606-8502, JAPAN.

Abstract

An inverse modeling to reproduce a variably saturated water flow in non-isothermal soil based on field observation is proposed. Since the water movement in the surface soil is significantly affected by the soil temperature, the governing equations system is composed of the mixed form Richards equation for the water movement and heat conduction equation for the thermal transport. To complete the water flow model of interest, unknown model parameters are determined with inverse technique. The major unknown parameter is the relative hydraulic conductivity (RHC) described as a free-form parameterized function which is a sequential piecewise cubic spline function and therefore can express the flexible functional form of the parameter. The inverse problem is defined as the minimization of errors between the observed and computed pressure heads to determine the coefficient values of the free-form function, and solved through a simulation-optimization method. To validate the water flow model developed, its practical application to in-situ soil is implemented. The results show that the functional form of RHC is successfully identified, and that both water movement and thermal transport models can produce the forward solutions which are good agreement with observed data for desorption period.

Key words: Parameter identification, Soil hydraulic properties, Free-form parameterization, Mixed-form Richards Equation, Heat conduction equation, Simulation-optimization method

1. INTRODUCTION

In sustainable agriculture, it is very important to optimize stable crop production and minimize pollution from agricultural chemicals and fertilizer with managing water efficiently. For this, an understanding of water movement through soil material is required. The mathematical model governing fluid flow in soil is represented by Richards equation (Richards, 1931). Generally, analytical solutions of Richards equation are not possible except under very restricting assumption due to the strong nonlinearity of the parameters involved. Instead, there have been many attempts to develop numerical methods (Hillel, 1998).

Because of the great advances of computer technology in recent years, numerical methods have increasingly attracted lots of attention of researchers and technical experts in many branches of science and engineering. The success of these numerical methods depends on the model structure given by governing equations system and parameter identification which is a critical step in modeling process.

Richards equation can be classified into three different forms: a pressure-head-based ($\psi$-based) form, a moisture-content-based ($\theta$-based) form, and a mixed form. The most commonly used numerical methods is associated with the $\psi$-based form of the equation because it can express all possible range of saturation while $\theta$-based form is not applicable for saturated zones where the hydraulic diffusivity takes infinite values. However, the disadvantages of $\psi$-based form are that it may require very small time and space intervals to achieve satisfactory mass balance (Hillel, 1998). Celia et al. (1990) presented an alternative model which pertains to the mixed form of Richards equation perfectly conserving mass balance. Since this model deals with only unsaturated zone, some extended models for saturated zone as well as unsaturated zone have been proposed (Takeuchi et al., 2008; Zadeh, 2011).

Richards equation expresses the water movement through soil derived from the gradient of hydraulic head, while water movement is also affected by soil temperature. Thus, this equation is not suitable to represent water
movement near surface soil where heat energy exchange is quite significant. As extended models, coupling models for simultaneous transfer of water and heat based on the work of Philip and de Vries (1957) have been proposed (e.g. Milly, 1982; Kondo and Saigusa, 1994; Fujiwara, 1995). While these models are related to the \( \psi \)-based form or \( \theta \)-based form of Richards equation coupled with heat transport equation, few coupling models of the mixed form of Richards equation and the heat transport equation has been studied.

On the other hand, the earlier works for parameter identification problem have commonly treated the well-defined models describing the soil hydraulic properties (i.e. the soil water retention curve and the unsaturated hydraulic conductivity) by a fixed-form function, such as the van Genuchten-Mualem model (Mualem, 1976; van Genuchten, 1980). Though the fixed-form functions make the inverse method relatively easy-to-handle due to the limited number of parameters to be determined, drawbacks are caused in employing this type of the function. Bitterlich et al. (2004) indicated the drawbacks as follows: (a) systematic errors in the fit of the retention function may propagate into the estimated conductivity function in cases where the selected fixed-form function is not flexible enough to represent the actual hydraulic properties, because the small number of parameters may be achieved in part by coupling the retention function with the conductivity function. Through common parameters, and (b) in the fixed-form function the degree of freedom has a global character in that parameter changes will affect the hydraulic functions over the entire pressure head or water content domain even if only one degree of freedom is modified. Hence, Bitterlich et al. (2004) proposed a free-form parameterization approach using piecewise polynomial functions as an alternative to the classic parametric approach. The validity of the approach proposed by Bitterlich et al. (2004) and its modified approach (Iden and Durner, 2007, 2008) are examined based on synthetic data sets and measurements through the multistep outflow or evaporation experiments in laboratory scale. Izumi et al. (2008, 2009) proposed a field-oriented approach for the inverse estimation of soil hydraulic properties based on the free-form parameterization because the laboratory experiments cannot be performed under fully natural conditions.

In the field experiments, the measurement of actual evapotranspiration is generally difficult. Zhang et al. (2010) embedded FAO approach (Allen et al., 1998) to calculate the potential evapotranspiration in the forward solution procedure of their inverse method to infer the soil hydraulic properties at the field scale. On the other hand, Izumi et al. (2009) avoided the difficulty in the estimation of the actual evapotranspiration by means of imposing the observed water content values as the Dirichlet boundary condition on the forward solution procedure, and thereby reduced the uncertainty in the identification of the unknown parameters.

The inverse methods proposed by Izumi et al. (2008, 2009) were associated with the \( \psi \)-based form and \( \theta \)-based form of Richards equation, and that by Izumi et al. (2010) was related to the mixed form of Richards equation under isothermal assumption. The purpose of this paper, thus, is to develop an inverse modeling for the mixed form of Richards equation in non-isothermal soil. Firstly, the governing equations system (forward problem, FP) is described, and the soil hydraulic properties which are model parameters included in the system are parameterized. The relative hydraulic conductivity (RHC) which is a major unknown parameter to be identified in this study is described by a free-form parameterized function which is a sequence of piecewise cubic spline functions over the whole effective saturation domain. For the representation of the soil water retention curve (SWRC), van Genuchten model (VG model) is employed due to being time-proven. Secondly, the inverse problem (IP) is defined as minimizing errors between the observed and computed values of the pressure head based on a simulation-optimization algorithm with the aid of the Levenberg-Marquardt method to determine the functional shape of RHC. Finally, the applicability of the inverse modeling developed is assessed through in-situ experiments in terms of reproducibility for observed water movement during the desorption process in that considering the hysteretic phenomenon is not needed.

2. GOVERNING EQUATIONS SYSTEM

2.1 Water Movement Model

For the water movement, the mixed form of Richards equation derived from the generalized Darcy’s law and the Boussinesq assumption (Hyakorn and Pinder, 1983) is employed to obtain the mass-conservative numerical solutions. Considering the dependency of density and viscosity of water on soil temperature to couple with thermal transport, the equation in one-dimensional vertical and saturated-unsaturated flow in which the liquid phase has considerable magnitude (i.e. neglecting the vapor fluxes) is described as follows:

\[
\phi \frac{\partial S_{w}}{\partial t} + WS_{s}S_{t} \frac{\partial \psi}{\partial t} = -\frac{\partial}{\partial z} \left( -K \frac{\partial h}{\partial z} + \frac{\rho_{r} - \rho_{l}}{\rho_{l}} \right) \tag{1}
\]

with

\[
W = \begin{cases} 
1 & (\psi \geq 0) \\
0 & (\psi < 0)
\end{cases} \tag{2}
\]

\[
S_{s} = \rho_{l} g (\beta_{s} + \phi \beta_{w}) \tag{3}
\]

\[
K = K_{r} (S_{s}) K_{r} (T_{r}) K_{s} \tag{4}
\]
\[ h = \frac{P}{\rho g} + z = \psi + z \]  \hspace{1cm} (5)

where \( \phi \) is the porosity, \( S_w \) the saturation, \( S_s \) the specific storage, \( \psi \) the pressure head, \( K \) the unsaturated hydraulic conductivity, \( h \) the hydraulic head, \( t \) the time, \( z \) the height defined as positive upward, \( \rho \) the water density at the soil temperature \( T_0 \), \( g \) the gravitational acceleration, \( \beta_s \) and \( \beta_w \) the compressibility coefficients of soil and water, respectively, \( K_r \) the relative hydraulic conductivity, \( K_t \) the correction-factor function of soil temperature, \( K_s \) the saturated hydraulic conductivity, \( S_e \) the effective saturation and \( p \) the water pressure.

2. 2 Thermal Transport Model

The heat flux due to the water movement in soil is smaller than the heat conduction by the solid soil and thus can be neglected. Accordingly, the heat conduction equation assuming a local thermal equilibrium between soil and water is employed for the thermal transport, and described based on Kondo and Saigusa (1994) as follows;

\[ \frac{\partial (C_b T)}{\partial t} = -\frac{\partial}{\partial z} \left( \lambda \frac{\partial T}{\partial z} \right) \]  \hspace{1cm} (6)

with

\[ C_b = (1-\phi)c_i + \theta e \]  \hspace{1cm} (7)

\[ \lambda = \lambda_0 + 0.5\beta_t \]  \hspace{1cm} (8)

where \( C_b \) is the volumetric heat capacity of soil, \( \theta \) the volumetric water content, \( c_i \) and \( e \) the volumetric heat capacity of soil particles (1.26 \times 10^6 [J/(m^3 K)]) and that of water (4.20 \times 10^3 [J/(m^3 K)]), respectively, and \( \lambda \) the thermal conductivity of soil expressed in terms of the reference thermal conductivity \( \lambda_0 \) and volumetric water content.

2. 3 Representation of Soil Hydraulic Properties

Soil hydraulic properties to be identified are the unsaturated hydraulic conductivity and soil water retention curve.

The unsaturated hydraulic conductivity is described as the product of three variables shown in Eq.(4). The correction-factor function of soil temperature and saturated hydraulic conductivity are represented as follows;

\[ K_t = \frac{\mu_t}{\mu_r} \]  \hspace{1cm} (9)

\[ K_s = \frac{\rho g \kappa}{\mu_t} \]  \hspace{1cm} (10)

where \( \mu_t \) and \( \mu_r \) are the dynamic viscosity coefficient at temperature \( T_t \) and \( T_r \), respectively, and \( \kappa \) the intrinsic permeability. Because the dynamic viscosity is the function of the temperature and the saturated hydraulic conductivity can be determined through laboratory experiments by definition, the representation of RHC function is needed.

To represent RHC, a sequential piecewise cubic spline function is employed. This is referred to as a free-form approach indicating that no a priori shape of RHC is assumed except for their monotonicity (Bitterlich et al., 2004). In the free-form approach, as shown in Fig.1, the effective saturation domain of interest is partitioned into \((I-1)\) subdomains with \( I \) nodes \((I\) denotes the degrees of freedom of the parameterization), and the function for \( K_i (S_e) \) over whole domain \([S_e (\theta_l), S_e (\theta_r)]\) is expressed by a summation of piecewise cubic spline interpolation functions which is locally defined over a confined subdomain bounded with two nodes. Thus, the function for \( K_i (S_e) \) with \( K_{ij} (S_e) \) defined over the \( i \)th subdomain \([S_{e,i}, S_{e,i+1}]\) is described as follows;

\[ K_i (S_e) = \sum_{j=1}^{I} K_{ij} (S_e) \]  \hspace{1cm} (11)

with

\[ K_{ij} (S_e) = \begin{cases} a_i + b_i (S_e - S_{e,i}) + c_i (S_e - S_{e,i})^3, & S_e \in [S_{e,i}, S_{e,i+1}] \\ 0, & S_e \notin [S_{e,i}, S_{e,i+1}] \end{cases} \]  \hspace{1cm} (12)

where \( a_i, b_i, c_i \) and \( d_i \) are coefficients in the cubic splines, and \( i (1 \leq i \leq I) \) a nodal number. Hereinafter, the values of \( K_i (S_e) \) at a node \( i \) are simply denoted by \( k_i \). Since, as well known, RHC monotonously increases with the increasing saturation, Eq.(11) must be identified so that the following constraints are satisfied.

\[ k_i \leq k_{i+1} \]  \hspace{1cm} (13)

SWRC is a constitutive relation between the volumet-
ric water content and pressure head. Various models have been proposed to represent SWRC. Among them, VG model has been more frequently used for the simulation of water movement in soil, and thus VG model is employed for the representation of SWRC in this study. Additionally, to account for the effect of soil temperature when calculating $\theta$ using VG model, the value of $\psi$ obtained from Eq.(1) is multiplied by two temperature correction coefficients. One is the ratio of the surface tensions at the soil temperature of interest and a reference temperature, and another is that of water density. This results in the following:

$$S_t(\psi_t) = \frac{1}{(1 + (\alpha_t \phi_t)^n)^{\mu_s}}$$

with

$$m_w = 1 - \frac{1}{n_w}$$

$$\psi_t = \frac{\sigma_s}{\sigma_t} \frac{\rho_t}{\rho_s} \psi$$

where $\theta_t$ is the residual water content, $\theta_s$ the saturated water content, $\alpha_t$, $m_w$, and $n_w$ the unknown parameters, $\psi_c$ the pressure head at a reference soil temperature $T_r$, and $\sigma_s$ and $\sigma_t$ the surface tension at soil temperature $T_0$ and a reference temperature $T_r$, respectively.

2. 4 Numerical Procedure

After discretization with the combination of the standard Galerkin finite element method for space and the finite difference method for time, Eqs.(1) and (6) are subjected to the following initial and boundary conditions and numerically solved with the iterative partitioned method;

$$\psi(z,0) = \psi_0(z) \quad \text{in } \Omega$$

$$T_s(z,0) = T_0(z) \quad \text{in } \Omega$$

$$\psi(z,t) = \bar{\psi}(z,t) \quad \text{on } \Gamma^w_0$$

$$T_s(z,t) = \bar{T}_s(z,t) \quad \text{on } \Gamma^h_0$$

$$-K_s K_r \frac{\partial h(z,t)}{\partial z} = \bar{q}_s(z,t) \quad \text{on } \Gamma^w_s$$

$$-\lambda \frac{\partial T_s(z,t)}{\partial z} = \bar{q}_s(z,t) \quad \text{on } \Gamma^h_s$$

where $\psi_0(z)$ and $T_0(z)$ are the initial value of the pressure head and soil temperature, respectively, $\Omega$ the space domain, $\bar{\psi}(z,t)$ and $\bar{T}_s(z,t)$ the values of the pressure head and soil temperature on the Dirichlet boundary, respectively, $\Gamma^w_s$ and $\Gamma^h_s$ the Dirichlet boundary for the water movement and heat transport, respectively, $\bar{q}_s(z,t)$ and $\bar{q}_h(z,t)$ the water and heat flux on Neumann boundary, respectively, and $\Gamma^w_n$ and $\Gamma^h_n$ the Neumann boundary for the water movement and heat transport, respectively.

3. PARAMETER IDENTIFICATION PROCEDURE

The unknown parameters to be identified are the soil hydraulic properties, i.e. RHC and SWRC, as described above. If the parameters can be determined with certainty in advance, they should be treated as known parameters because the number of identified parameters should be kept as small as possible. RHC, in general, cannot be determined by the direct measurement while SWRC can be obtained from the time-series data of the pressure head and volumetric water content with relative ease. Hence, RHC is treated as the unknown parameter and is identified with use of inverse technique.

3.1 Inverse Problem (IP)

To solve IP is to optimally decide the unknown nodal values, $k = \{k_i, 1 \leq i \leq L\}$, which minimize the objective function defined as the total least squares error integrated over space and time between the solution of FP $(\psi_{\text{sol}}(k))$ and the observed data $(\psi_{\text{obs}})$, and thereby IP is defined as follows;

$$J(k_{\text{opt}}) = \min J(k), \quad k_{\text{opt}}, k \in K_{ad}$$

with

$$J(k) = \frac{1}{2} \sum_{i=1}^{L} (f_i(k))^2$$

$$f_i(k) = w_i \left( \psi_{\text{sol}}(k) - \psi_{\text{obs}} \right)$$

where $J(k)$ is the objective function, $k_{\text{opt}}$ the set of optimal solutions, $K_{ad}$ an admissible set of $k_{\text{opt}}$ and $k$, $L$ the total number of observed data available in space and time, and $w_i$ (normally, taken as unity) a weighting factor. The constant 1/2 in Eq.(24) is an additive constant to cancel the differential coefficient obtained from differentiating the sum of squares.

The process of solving IP is shown in Fig.2. The decision variables, $k$, are iteratively modified or updated while step by step solving FP with the assumed or previously estimated values. In this respect, identification of RHC function (Eq.(11)) requires a sort of simulation-optimization technique.

3.2 Optimization Algorithm

Levenberg-Marquardt method, which is a modified Gauss-Newton method combining the Gauss-Newton method and the gradient method, is employed for the optimization algorithm to search for the set of optimal solution with following search sequence through the iteration
\(k^{(r+1)} = k^{(r)} + \Delta k^{(r)}\)  \hspace{1cm} (26)

with

\[
\Delta k^{(r)} = -\left( H^{(r)} + \eta I \right)^{-1} \nabla f^{(r)}
\]

\(H^{(r)} = \left[ \frac{\partial^2 f^{(r)}}{\partial k_i \partial k_j} \right] \)

where \(r\) is an iteration number, \(\eta\) a coefficient which controls search strategy between the Gauss-Newton and the steepest descent direction, and \(I\) the \(I \times I\) unit matrix.

### 4. VALIDATION

Compared with the inverse modeling under isothermal assumption proposed by Izumi et al. (2010), the applicability of this inverse modeling is assessed in terms of reproducibility for observed water movement in test soil. The time-series observed data is obtained through \textit{in-situ} experiments (sandy soil) in Matsuyama, Ehime Prefecture.

#### 4.1 Field Observation and Computational Domain

The observation system and computational domain for parameter identification are illustrated in Fig. 3.

The observation system consists of three sets of instruments—tensiometer (UIZ-SMT), soil moisture probe (UIZ-SM-2X) and thermometer (TMC20-HD)—which automatically records time-series data of the pressure head, volumetric water content and soil temperature at intervals of 10 minutes. To reduce the influence of direct solar radiation on the observed pressure heads, the sensors of tensiometers attached above ground are shielded by white-colored box with slits for ventilating air. Each set of instruments is buried at three different depths in test soil: 

-10 cm, -20 cm and -30 cm. A pair of pressure head and volumetric water content data at the same depth is used to determine SWRC. The observed values of the pressure head at -20 cm deep is used to determine the functional shape of RHC (Eq.(11)) and to assess the fitness of estimated RHC function in the simulation-optimization runs. The pressure head and soil temperature observed at both top and bottom (-10 cm and -30 cm deep) of test soil are utilized as Dirichlet boundary values for solving FP to intentionally avoid measuring flux like evapotranspiration whose measurement is generally difficult in the field experiments. The time-series soil temperatures at -20 cm deep are used as the benchmark data for confirming reproducibility of forward solution.

The computational domain therefore is the surface soil of 20 cm thick between -10 cm and -30 cm deep, where the saturation and soil temperature vary significantly with meteorological conditions, and is divided into four equal elements with five nodes for numerical computations.

Like in most earlier works on parameter identification, a one-way process of desorption is considered for estimation of the soil hydraulic properties to exclude the hysteretic phenomenon in this study. Hence, the data series observed during no-rainfall or desorption period—November 25 to December 2, 2010—and used for parameter identification and validation of the inverse modeling, is shown in Fig. 4.

#### 4.2 Identification of SWRC

SWRC is defined as the approximate curve for the scatter plots relating volumetric water content to pressure head, and should be determined before estimating RHC in this inverse modeling. To obtain \(\theta - \psi\) relations at a refer-
ence soil temperature expressed by \( \text{Eq.(14)} \), van Genuchten et al. (1991) proposed the RETC program which is an estimation method for unknown parameters included in SWRC function with use of the nonlinear least-squares approach. Generally, it is very important to input an adequate estimate of the initial value of unknown parameters for such nonlinear fitting. Since the RETC program does not have the function of automatically estimating the initial parameters, Seki (2007) developed the alternative algorithm called SWRC fit. This nonlinear fitting software not only can automatically determine the initial estimate of unknown parameters, but also can accomplish the parameter identification for the five SWRC models including VG model. In this study, therefore, SWRC fit is employed to identify the reference \( \theta - \psi \) relations.

For the implementation, observed data of \( \theta \) and \( \psi \) are corrected for a reference temperature based on the soil temperature at the same depth, using the coefficient of thermal expansion and Eq.(16), respectively. Additionally, the saturated water content \( \theta_s \) is defined as the value of \( \theta \) in case that \( \psi \) is equal to zero while the residual water content \( \theta_r \) as the minimum value in observed data of \( \theta \). Thus, \( \theta_s \) and \( \theta_r \) are determined to be 0.30 and 0.05 from all observed data series including them shown in Fig.4, respectively. As the reference soil temperature, the average value of observed data is adopted which equals to 12.4 degrees Celsius.

Conclusively, a best fitting curve for \( \theta - \psi \) relation at the reference soil temperature is obtained with the value of \( \alpha \) and \( n_\psi \) being 28.9 m\(^{-1}\) and 1.43, respectively, as shown in Fig.5.

4.3 Identification of RHC
Firstly, the number of identified unknown nodal values is determined. In the free form approach, the large number of nodal values can be set for the functional form of RHC to have high flexibility or degree of freedom. However, the large number of nodal values cause the increasing uncertainty of identified nodal values, making IP difficult to solve. For this, the number of identified nodal values should be kept to the minimum in general. Since the optimal number of nodal values is approximately from seven to nine according to Bitterlich et al. (2004) and Iden and Durner (2007), the number of unknown nodal values is set to be 10 in this study.

Secondly, the manner of division on the definition domain (effective saturation domain) is determined. The functional shape of RHC generally has steeper gradient near the saturated zone. In the context of accuracy, the range with steeper gradient should be divided into finer subdomain. Therefore the effective saturation range from 0.8 to 1.0 where RHC function rapidly changes is partitioned into sub-range in the manner of geometric series.

Thirdly, the reference thermal conductivity \( K_0 \) is determined based on Kondo and Saigusa (1994).

Eventually, RHC function is identified as shown in Fig.6 after the saturated hydraulic conductivity at the reference temperature is determined to be \( 5.18 \times 10^{-5} \) m/s through laboratory experiment.
4. Reproducibility of Calibrated Forward Simulation Model for Variably Saturated Water Flow

The applicability of the inverse modeling presently developed is assessed in terms of reproducibility for the time-series observed data by calibrated forward simulation model through comparing the observed and computed values. The result of reproducibility for water movement and heat transport are shown in Figs. 7 and 8, respectively. In the lower half of Fig. 7, the absolute errors \( |\psi_{\text{obs}} - \psi_{\text{com}}| \) are also shown to demonstrate the time-varying difference in solution reproducibility. From the results, it is found that both time-varying pressure head and soil temperature are reproduced with high accuracy.

For the purpose of comparison, the preceding inverse modeling proposed by Izumi et al. (2010), which did not consider the influence of soil temperature on water movement, is applied to the same observed data. In Figs. 9 and 10, the results for identification of RHC and the reproducibility of forward solution for water movement are shown in a similar trend of Figs. 6 and 7, respectively. Hereinafter, the proposed model in this paper is referred to as the thermal model and the model proposed by Izumi et al. (2010) is referred to as the isothermal model. In addition, the values of the objective function \( J(k) \) for both models are summarized in Table 1 to supply superiority of the thermal model with a quantitative underpinning. The comparison results indicate that consideration of temperature-dependency is beneficial to simulate the water flow near the soil surface, and thus it concludes that the inverse model proposed provide a high performance for the modeling of water flow.

5. CONCLUSION

An inverse modeling for variably saturated and non-isothermal water flow is developed. To consider the
water movement depending on the soil temperature, a couple of mixed form of Richards equation with heat conduction equation is employed as the governing equation which describes the forward problem. The soil hydraulic properties which are the model parameters included in the governing equation are the soil water retention curve and unsaturated (or relative) hydraulic conductivity. Since the former is determined through experiments with relative ease, the latter is unknown parameter in this study. For the functional form of the relative hydraulic conductivity, the tree-form approach is employed. The inverse problem to identify the relative hydraulic conductivity function is defined and solved for the practical field experiments with in-situ soil near the surface soil to evaluate the performance of proposed method. From the examination and comparison of the preceding method which did not consider the influence of soil temperature on the water movement, the advantage and appropriateness of this method is shown and the importance of consideration for the temperature-dependency in the water movement near the soil surface is emphasize.

REFERENCES

[Received 2011. 8. 8. Accepted 2012. 10. 31] (Questions and/or discussions on this paper for public debate will be accepted before 2013. 6. 24.)