

Experiments on the Control of Salinity and Sodicity in Surface : Irrigated Fields in the Upper Yellow River Valley (II)

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Experiments on the Control of Salinity and Sodicity in Surface-Irrigated Fields in the Upper Yellow River Valley (II)

Observation System and Calibration

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An experimental field was established in Pingbu village extending along the Yellow River (Baiyin, Gansu, China) to conduct experiments on the control of salinity and sodicity in surface-irrigated fields, and an observation system for measuring meteorological and hydrological elements relevant to soil salinization and alkalization was installed in it. The outline of the observation system is shown; principles of measuring such elements are examined and their limitations are pointed out in case of necessity. Furthermore, careful considerations are given to the system so that the system could measure such elements as accurately as possible.

KEY WORDS: biological reclamation, irrigation, salinity, soil physical reclamation, sodicity, TDR

INTRODUCTION

For irrigation agriculture to succeed, salinity and sodicity levels in soils and irrigation water must be controlled. To conduct experiments on the development of effective, soil physical and biological measures for controlling salinity and sodicity in surface-irrigated fields, an experimental field was established in Pingbu, Gansu, China (Wang *et al.*, this issue).

Since the electrical conductivity (EC) of soil solution is linearly related to the total salt concentration, soil salinity can be monitored continuously and automatically in the field by measuring EC with TDR (time domain reflectometry) probes (Wang *et al.*, 2006). As for sodicity, however, it is difficult to measure it (SAR or ESP) automatically in the field (Wang *et al.*, this issue). Therefore, water extracts of soils are used for measuring soil sodicity in the laboratory. In general, however, the saturation extract or dilution extracts such as the extract of a soil sample at 1: x soil–water ratio ($x=1\sim5$) is used (Wang *et al.*, this issue; Yasutake *et al.*, this issue).

Salts in soil are transported by soil water, and therefore soil water movement has to be monitored in order to analyze the salt behavior in soil. Thus, it is essential to observe precipitation, soil moisture, groundwater level, and evaporation in the field for conducting experiments on soil salinization processes.

This paper gives the outline of the observation system installed at the experimental field. The principle of the TDR technique and methods of measuring evaporation in the field are also examined, and the results of calibrating this system are shown, which calibration was made so that this system could be used for measuring the variables relevant to this subject as accurately as possible.

INSTRUMENTS

The experimental field established at Jiangtan section of Pingbu village in Jingyuan, Baiyin, Gansu (N36° 25.5', E104° 25.4', 1461 m ASL) is located in an alluvial valley of the Yellow River, and a cliff of the yellow loess about 150 m high rises about 500 m southeast of the field. **Fig. 1** shows a ground plan of the field, the area being about one mu (Chinese unit of area, 667 m²). Although part of the northwest edge shares with the residential lot of the owner of this field, this field is bordered on fields or a narrow road to the other directions.

The arrangement of instruments is shown in **Fig. 1** and the general description of the instruments is listed in **Table 1**. The water diverted from the Yellow River flows through the canal excavated along the road at all times in the growing season and the part of the experimental field close to the canal is severely salinized. Thus, three points for measuring soil moisture and salinity were arranged so that one (P1) was in the severely salinized

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Table 1. Meteorological and hydrological instruments installed in the experimental field

Variable	Instrument	Sampling & averaging Δt	Height or depth (m)
Air temperature & humidity	VAISALA, HMP50	15 sec & 10 min*	1.8, 4.3
Net radiation	CAMPBELL, Q*7	As above	2.8
Solar radiation	CAMPBELL, LI200X	As above	2.8
Wind speed & direction	YOUNG, CYG5103	As above	6.0
Ground heat flux	REBS, HFT-3.1	As above	-0.05, -0.15
Precipitation	CAMPBELL, TE525	10 min (cumulative)*	-
Soil moisture & EC	CAMPBELL, CS610	30 min (momentary)	-10, -20, -40, -60 (P2)
Soil temperature	CAMPBELL, MO 107	As above	-10, -20, -40
Groundwater level	CAMPBELL, CS420	As above	-

* On and after 10 Nov. 2007, time period 10 min was changed to 30 min.

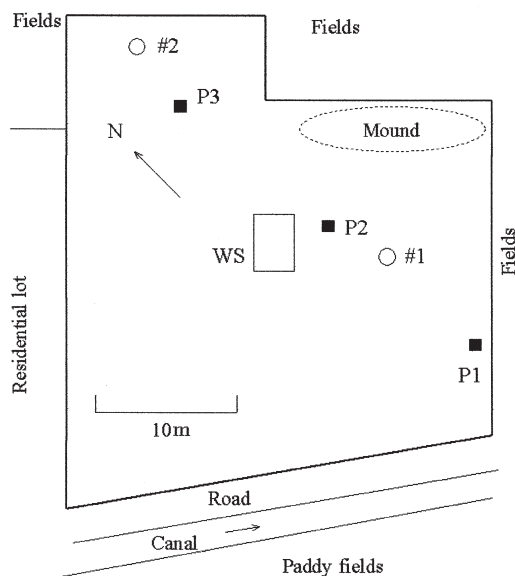


Fig. 1. A ground plan of the Pingbu experimental field. WS: weather station, ○: well, ■: TDR.

part and the others (P2 and P3) were moderately and slightly salinized parts in the field, respectively (**Fig. 1**).

Volumetric soil water content (θ) and bulk soil electric conductivity (EC_a) are measured using TDR probes at P1, P2 and P3. Two tube wells were dug in the field (#1 and #2). The well #1 is for not only measuring the groundwater level but also pumping groundwater out, the diameter being 11 cm and the depth to the bottom end of the perforated pipe being 210 cm (measurable depth limit = -193 cm). On the other hand, the well #2 is for observing the groundwater level, the diameter being 8.5 cm and the depth being 130 cm (measurable depth limit = -110 cm). In this field, depth to the water table is smaller than 1 m in the growing season when the canal is filled with irrigation water, though it becomes larger than 1 m in winter when the canal is dry and precipitation is low (Wang *et al.*, this issue).

A meteorological pole 6 m high was raised in an meteorological observation field enclosed with a chain-link fence (3×5 m, WS in **Fig. 1**). Wind speed and direction, air temperature and humidity, solar radiation, and net radiation are measured with sensors mounted on the pole (**Table 1**). A raingage and ground heat flow meters are also set in WS, while soil thermometers were installed at P2.

PHYSICAL PROPERTIES OF THE SOIL

Physical properties of the soil in the Pingbu experimental field (P1), which is an alluvial soil, are listed in **Table 2**. Those of the soil in the experimental field established at Togtoh, Inner Mongolia (Wang *et al.*, this issue) are also shown for reference. **Fig. 2** shows the matric potential (suction)–water content relation for the Pingbu soil. The curve is a van Genuchten equation fit to the measurements (Jury and Horton, 2004). **Fig. 3** shows the matric potential (suction)–hydraulic conductivity relation for the Pingbu soil obtained by the van Genuchten parametric model (Jury and Horton, 2004).

It is obvious from **Table 2** that 89.1 % of the Pingbu

Table 2. Physical properties of the soil in the Pingbu field (P1). Those of the soil in the Togtoh field also are shown for reference

Particle size (mm)	Pingbu (P1)	Togtoh
	% by weight	
<0.002	14.9	10.3
0.002 ~ 0.075	74.2	67.5
0.075 ~ 0.25	10.9	17.3
0.25 ~ 0.85	0	0.2
0.85 ~ 2	0	4.7
Particle density ($g\ cm^{-3}$)	2.71	2.70
Dry density ($g\ cm^{-3}$)	1.60	1.40
Porosity (-)	0.40	0.50
Saturated hydraulic conductivity ($cm\ s^{-1}$)	3.13×10^{-5}	1.29×10^{-4}

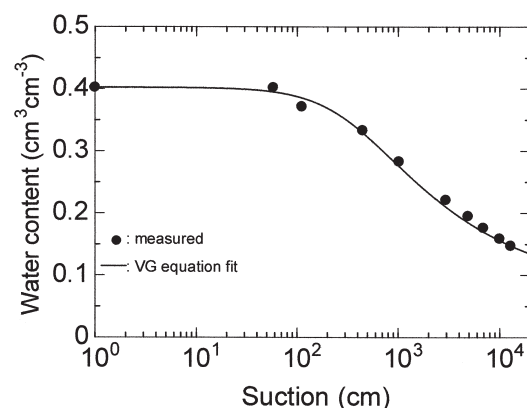


Fig. 2. Matric potential (suction)–water content relation for the soil in this field (P1). The curve is a van Genuchten equation fit to the measurements ($\theta_s=0.066$, $\theta_r=0.40$, $\alpha=0.028$, $N=1.40$).

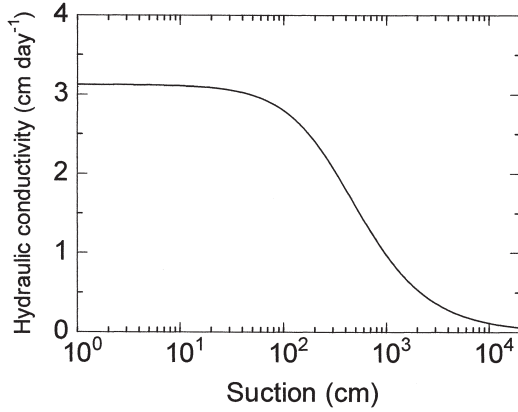


Fig. 3. Matric potential (suction)–hydraulic conductivity relation for the soil in this field (P1) obtained by the van Genuchten parametric model ($\theta_s=0.066$, $\theta_r=0.40$, $\alpha=0.028$, $N=1.40$, $K_s=3.13 \times 10^{-5}$).

soil is smaller in diameter than $0.05 \mu\text{m}$, while the portion is 77.8% for the Togtoh soil. Furthermore, the saturated hydraulic conductivity of the Pingbu soil is about an order of magnitude smaller than that of the Togtoh soil. These properties suggest that the Pingbu soil is more susceptible to salinization than the Togtoh soil.

TIME DOMAIN REFLECTOMETRY (TDR)

Principle

Time domain reflectometry (TDR) is used for simultaneous measurement of volumetric soil water content (θ) and bulk soil electric conductivity (EC_a) in this field. The measurement of θ is based on measuring the travel time of an electromagnetic pulse lunched along a transmission line embedded in the soil, while the measurement of EC_a is based on measuring the impedance of the soil.

The main components of a TDR system are a cable tester, a coaxial connection cable, and the TDR probe. The cable tester sends out a high-frequency electromagnetic pulse (~ 1 GHz), which travels down the coaxial cable and enters the TDR probe. At both ends of the probe, the pulse is partly reflected back, from which the travel time of the pulse along the probe can be detected.

If K' is the real part of the complex dielectric constant K^* , or

$$K^* = K' - i(K'' + \sigma_{dc} / \omega \epsilon_0) \quad (1)$$

where i =imaginary unit, ω =frequency of alternating current, σ_{dc} =direct current conductivity, ϵ_0 =permittivity of free space (Takahashi, 1970), the speed (v) of an electromagnetic wave in nonmagnetic material such as soil and solution is approximately given as

$$v = \frac{c}{K'^{1/2}} \quad (2)$$

because $K' \gg (K'' + \sigma_{dc} / \omega \epsilon_0)$ (Yanuka *et al.*, 1988).

The soil comprises a volume fraction of soil solids, water, and air. Representative values of $K_a = |K^*|$ are

80.36 for liquid water, 3–5 for soil materials, 6–8 for organic matter, and 1 for air at a frequency of 1 GHz and a temperature of 20 °C (Vogeler *et al.*, 2005). Thus, K_a ($\sim K'$) of a soil is dominated by the content of water because of its high K_a value. The relation between θ and K_a can be determined experimentally (Eq.13).

A simple model of the TDR system is used to explain how to measure the impedance of a soil using TDR. The model is described as a coaxial transmission line consisting of two conductors, with a power source (i.e., a cable tester) at one end (say, left) and a load (i.e., the soil sample) having impedance Z_L at the other end (right).

The basic equations for a transmission line can be written as (Takahashi, 1970):

$$\begin{aligned} -i\omega CE &= \frac{dI}{dx} \\ -i\omega LI &= \frac{dE}{dx} \end{aligned} \quad (3)$$

where C =capacitance per length, L =inductance per length, E =the difference in electric potential between the two conductors, I =electric current, x =coordinate along the line (positive rightward). The time dependence of I and E expressed by $e^{i\omega t}$ is omitted in these expressions.

The solutions to Eq.(3) are given by

$$\begin{aligned} E &= E_+ e^{-\gamma x} + E_- e^{\gamma x} \\ I &= I_+ e^{-\gamma x} + I_- e^{\gamma x} \end{aligned} \quad (4)$$

where

$$\gamma = i\omega \sqrt{LC}$$

and subscripts + and – indicate that the quantities are related to the wave propagating in the right and left direction, respectively. Therefore, characteristic impedance Z_0 is written as

$$Z_0 = \frac{E_+}{I_+} = -\frac{E_-}{I_-} = \sqrt{\frac{L}{C}} \quad (5)$$

If the origin of the coordinate x is set at the point of the load having impedance Z_L or the right end of the coaxial transmission line, the following relation is obtained using Eqs.(4) and (5).

$$Z_L = \frac{E(x=0)}{I(x=0)} = \frac{E_+ + E_-}{I_+ + I_-} = \frac{I_- - I_+}{I_+ + I_-} Z_0 \quad (6)$$

If the reflection coefficient at the right end of the transmission line is designated as ρ , then using Eqs. (5) and (6) we get

$$\rho = \frac{E_-}{E_+} = -\frac{I_-}{I_+} = \frac{Z_L - Z_0}{Z_L + Z_0} \quad (7)$$

Since the characteristic impedance Z_0 is an intrinsic value of the TDR system, if ρ can be measured using TDR, we can get Z_L from Eq.(7).

The impedance of a soil is, in general, indicated by a complex number. However, because Z_0 is a real number (Eq.5), Z_L is also regarded as a real number as is obvious from Eq. (6). This is, of course, a result of applying the

simple model to the TDR system, but the result is approximately true. Since the real part of impedance means the resistance, the TDR-measured impedance Z_L can be converted to EC_a as

$$EC_a = fK_G Z_L^{-1} \quad (8)$$

where f is a correction coefficient for temperatures (T °C) other than 25 °C and expressed as (Dirksen, 1999):

$$f = \frac{1}{1 - 0.0191(T - 25)} \quad (9)$$

K_G is the probe geometry constant and experimentally determined one by one (m^{-1}).

In practice, the reflection coefficient ρ is estimated with the "thin sample" equation derived by Giese and Tiemann (Topp *et al.*, 1988) based on the waveform analysis of the TDR trace (Dirksen, 1999).

If ρ is too small, or Z_L approaches extremely to Z_0 due to high soil salinity, the TDR system cannot get the information on the electromagnetic property of the soil. Furthermore, the relation between EC_a and Z_L^{-1} (Eq.8) is based on the assumption that the relation is intrinsically linear. It may be true at least for the electrical conductivity of the soil water (EC_w) up to 8~20 dS m^{-1} , but with higher conductivity, the relation becomes nonlinear (Vogeler *et al.*, 2005).

The degree of nonlinearity is influenced by the characteristics of the TDR probe, with three-wire sensors being superior to two-wire sensors (Vogeler *et al.*, 2005). In this experiment, the TDR probe consisting of three rods of length 30 cm, diameter 0.48 m, and spacing between rods of 2.2 cm was used and inserted horizontally into the soil (**Table 1**).

Models of electrical conduction in soil

Simultaneous measurements of θ and EC_a can be taken with a single TDR probe. However, it would be desirable to know the soil solute concentration in the soil water, or EC_w over the entire range of the field water contents and to obtain this information directly in the field (Rhoades, 1982). Various approaches to this issue exist.

Wang *et al.* (2005) adopted a model of electrical conduction in soil suggested by Rhoades *et al.* (1976) for analyzing soil salinization processes in the Togtoh experimental field, which gives the following relationship:

$$EC_a = EC_w \theta T(\theta) + EC_s \quad (10)$$

where

$$T(\theta) = \alpha\theta + \beta \quad (11)$$

is an empirical relation expressing a transmission coefficient (≤ 1) and α and β are soil-dependent constants, and EC_s is the electrical conductivity of the solid phase of a soil due primarily to exchangeable cations adsorbed

on clay minerals.

In Eq.(10), the relation between EC_a and EC_w is assumed to be linear for a fixed water content. However, at low EC_w under 1~5 dS m^{-1} , a nonlinearity in the EC_a - EC_w relation has been observed in some soils. Thus, if the linearity in the EC_a - Z_L^{-1} relation (Eq.9) is also taken into consideration, the TDR technique should be applied to soil solutions with a range of EC_w from about 1 to 20 dS m^{-1} (Vogeler *et al.*, 2005). If we identify the three parameters EC_s , α and β , EC_w can be estimated from the measurements of EC_a and θ made using TDR from Eqs. (10) and (11).

Wang *et al.* (2005) assumed that the relation between EC_w and θ could be written as

$$EC_w = EC_{SAT} \left(\frac{\varepsilon}{\theta} \right)^n \quad (12)$$

where EC_{SAT} = electrical conductivity of the extract of a saturated soil (Kobayashi *et al.*, 2006), ε = porosity of the soil, n = a parameter to be determined by experiment. When precipitation-dissolution reactions can be neglected, we can set $n=1$, while in general, $0 < n \leq 1$. Thus, EC_{SAT} is estimated from EC_w using Eq.(12) if n is determined by experiment.

Calibration

Soil samples were collected in the experimental field during a period of 12~13 Aug. 2007, when depth to the water table was about 50 cm and the EC of the groundwater was about 4 dS m^{-1} . However, the EC_{SAT} of the surface soil exceeded 10 dS m^{-1} , which means that salts were severely accumulated in the uppermost soil layers. Therefore, the calibration of the model of electrical conduction in soil (Eqs. 10 and 11), in which EC_w should be increased step by step from about 1 to 20 dS m^{-1} (Wang *et al.*, 2005) could not be practiced in the laboratory.

The second chance to take soil samples came on 19 Sep. 2007, when the EC_{SAT} of the surface soil collected around P3 was 4.5 dS m^{-1} (**Fig. 4**), while the EC of the groundwater measured at the well #2 was 2.2 dS m^{-1} and depth to the water table was 67 cm. The soil sample collected in the field seems to have been too much salinized to use for identifying the model parameters. Therefore, in order to identify the parameters, it seems necessary to devise an experimental technique with which the soil sample is desalted.

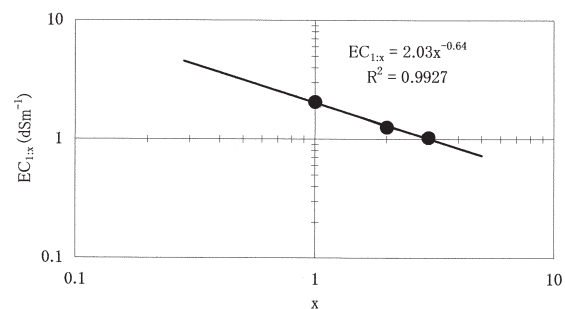


Fig. 4. Relationship between EC and x for the soil sampled around P3 on 19 Sep. 2007. The left and right ends of the regression line segment corresponds to the saturation or $x=0.28$ and $x=5$, respectively.

Thus, for the present, the results obtained for the soil of the Togtoh experimental field are given as the alternatives (Wang *et al.*, 2006). They are not expected too much different from the ones for the Pingbu soil, because both soils are alluvial soils originated from the Yellow River.

$$\theta = 1.65 \times 10^{-5} K_a^3 - 9.91 \times 10^{-4} K_a^2 + 3.32 \times 10^{-2} K_a - 4.8 \times 10^{-2} \quad (13)$$

$$EC_a = EC_w \theta (1.513 \theta - 0.074) + 0.08 \quad (\text{dS m}^{-1}) \quad (14)$$

$$(0.15 < \theta < 0.4, 0.15 < EC_a < 1.5 \text{ dS m}^{-1})$$

where every value of EC was normalized to a temperature of 25 °C.

EVAOPORATION MEASUREMENT

Methods

Meteorological and hydrological variables that are indispensable for monitoring water movement in the soil profile are precipitation and evaporation. In this experimental field, evaporation is measured by the Bowen ratio method except for the winter days when the surface soil freezes and/or thaws during the course of a day (Kaneko *et al.*, 2006). The aerodynamic method is used for such exceptional days.

Bowen ratio method

The energy balance of the soil surface is expressed as

$$R_N - G = H + \lambda E \quad (15)$$

where R_N =net radiation, G =ground heat flux, H =sensible heat flux, E =evaporation rate, λ =latent heat of vaporization. Eq.(15) can be rewritten in the form

$$\lambda E = \frac{(R_N - G)}{1 + \beta} \quad (16)$$

where $\beta \equiv H / \lambda E$ is the Bowen ratio.

Assuming that the transfer coefficients of heat and water vapor in the surface air layer are equal, the Bowen ratio β can be estimated from an approximate equation

$$\beta = \gamma \frac{T_2 - T_1}{e_2 - e_1} \quad (17)$$

where T_i ($i=1, 2$) = air temperature at height h_i , e_i =water vapor pressure at height h_i , γ = psychrometer constant.

Aerodynamic method

The aerodynamic method relies on the existence of relations between fluxes and gradients of the forms described below (Kaneko *et al.*, 2006).

$$\text{Momentum: } \tau = \rho u_*^2 = K_M \rho \frac{\Delta u}{\Delta z} \quad (18)$$

$$\text{Water vapor: } E = -K_V \frac{\Delta \rho_v}{\Delta z} \quad (19)$$

where ρ = density of air ($\approx 1.06 \text{ kg m}^{-3}$ at 1500 m ASL), $\Delta u / \Delta z$ =vertical gradient of wind speed, $\Delta \rho_v / \Delta z$ =vertical

gradient of water vapor density, u_* =friction velocity, K_M =eddy diffusivity of momentum, K_V =eddy diffusivity of water vapor.

If it is assumed that $K_M = K_V$, the combination of Eqs. (18) and (19) yields

$$E = - \frac{\Delta \rho_v}{\Delta u} u_*^2 = -\rho \frac{\Delta q}{\Delta u} u_*^2 \quad (20)$$

where q =specific humidity.

Since the surface is bare of plants in winter, when this method is used, the vertical gradient of wind speed can be written as

$$\frac{\Delta u}{\Delta z} = \frac{u_*}{\kappa z} \phi_M \quad (21)$$

where κ = von Karman constant (≈ 0.4), ϕ_M = dimensionless stability function. However, in this observation system, wind speed is measured at only one height, so the function ϕ_M cannot be evaluated. Therefore it is assumed that the static stability is neutral, or $\phi_M = 1$, which does not seem too much different from the actual daytime conditions over the frozen soil. Then integrating Eq.(21) gives

$$u = \frac{u_*}{\kappa} \ln \frac{z}{z_0} \quad (22)$$

where z_0 = roughness length. Thus, if the roughness length is determined, substituting Eqs.(23) and (24)

$$u_* = \frac{\kappa u}{\ln(z/z_0)} \quad (23)$$

$$\Delta u = u(z_2) - u(z_1) = \frac{u_*}{\kappa} \ln \frac{z_2}{z_1} \quad (24)$$

in Eq.(20) gives

$$E = -\rho [q(z_2) - q(z_1)] u \frac{\kappa^2}{\ln(z/z_0) \ln(z_2/z_1)} \quad (25)$$

where z = wind speed measurement height, z_1, z_2 : humidity measurement heights.

Calibration

Bowen ratio method

Although psychrometer constant, $\gamma \equiv C_p p / 0.622 \gamma$, where C_p =specific heat at constant pressure and p =air pressure, is proportional to p , air pressure is not measured in the experimental field (**Table 2**). Therefore, a global average of 846 hPa at 1500 m ASL is used. So a typical value for γ is 0.55 hPa K⁻¹.

Since air temperature and humidity are measured at only two heights in this field (**Table 2**), the accuracy of the measurements of T_i and e_i or q_i ($i=1, 2$) is essential to estimate the Bowen ratio accurately using Eq.(17). Therefore, before installing the sensors on the pole, a comparative measurement of air temperature and humidity with the two thermo-hygrometers mounted on the same arm was carried out in the experimental field for a full day 19–20 Sep. 2007. The results obtained are as follows:

$$\begin{aligned} \text{Air temperature (}^\circ\text{C): } T_2(\text{corrected}) \\ = 1.006 T_2(\text{measured}) - 0.26 \end{aligned} \quad (26)$$

$$\begin{aligned} \text{Relative humidity (\%): } RH_2(\text{corrected}) \\ = 1.010 RH_2(\text{measured}) - 0.33 \end{aligned} \quad (27)$$

Here the sensors installed at 4.3 m and 1.8 m AGL were numbered 1 and 2, respectively, and T_2 and RH_2 indicate the measurements made with the sensors designated No. 2. The sensors designated No. 1 were regarded as the standard ones (**Fig. 5**).

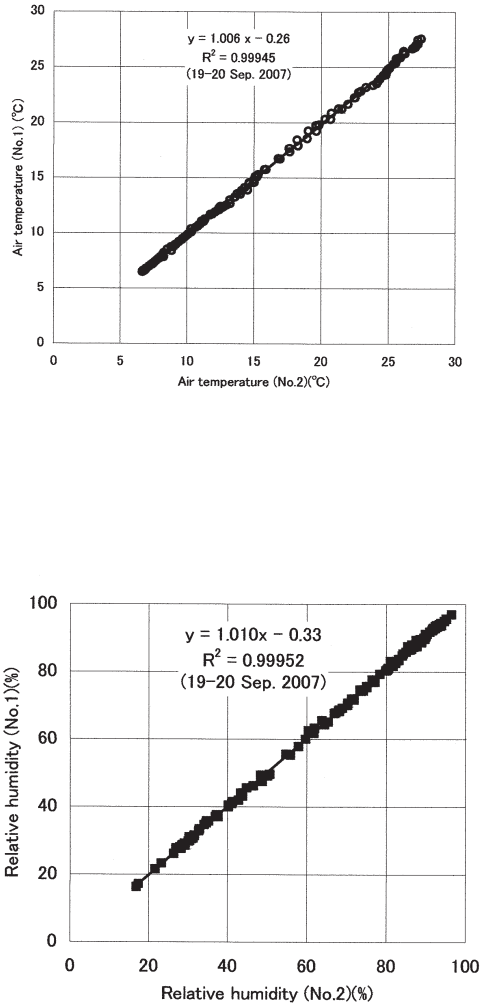


Fig. 5. Results of comparative measurements of (upper) air temperature and (lower) relative humidity with the two thermo-hygrometers mounted on the same arm.

Aerodynamic method

Eq.(25) is rewritten as

$$E = Bu\rho [q(z_1) - q(z_2)] \quad (28)$$

where

$$B \equiv \frac{\kappa^2}{\ln(z/z_0)\ln(z_2/z_1)} \quad (29)$$

In this observation system, $z=6.0$ m, $z_1=1.8$ m, and $z_2=4.3$ m (**Table 2**), so B is a function of only roughness length z_0 . A possible range of z_0 of the field without vegetation in winter may be from 10^{-4} m to 10^{-2} m. Thus, the value for B is estimated to range from about 2×10^{-2} to 3×10^{-2} .

When the aerodynamic method is used to estimate the evaporation from the soil surface in winter, the value

for z_0 should be identified in advance by comparing measurements of evaporation made by the aerodynamic method with those obtained by the Bowen ratio method in the field without vegetation during a unfrozen period.

CONCLUDING REMARKS

The observation system installed in the Pingbu experimental field for measuring meteorological and hydrological elements relevant to salinization and alkalinization of soils was described in detail. Principles of measuring such elements were examined and their limitations also were pointed out in case of necessity. Furthermore, careful consideration was given to the system so that the system could measure such elements as accurately as possible. A comparison of the physical properties of the soil in this field with those in the Togtoh experimental field suggested that the Pingbu soil is more susceptible to salinization than the Togtoh soil is.

The observation system, however, cannot give us sufficient information on the spatial variations in salinity in the field. Furthermore, elements relevant to alkalization and biological reclamation of surface-irrigated fields (Wang *et al.*, this issue) cannot be observed with this system. Therefore, in order to conduct these experiments, it is necessary to make intensive observations frequently in and around the field.

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