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Estimation of real and predicted infiltration recharge for the irrigated area in southern Ukraine

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Abstract Investigations were conducted to estimate real and predicted infiltration recharge rates within the Dnieper-Molochnaya interfluve and the Steppe Crimea, Ukraine. The recharge was used to predict the impact of present and planned irrigation systems on the environment such as underflooding, salinization, and drinking groundwater quality deterioration. Current infiltration recharge was estimated for 45 000 km^2 with a 4-km discretization during flow model identification. Infiltration recharge values and their uncertainties, which were obtained using direct or indirect measurements and calculations for several different sites, permitted determination only of assessed values. Predicted infiltration recharge and its concentrations were determined using vertical water and salt transport models reflecting the lithology and salinization of unsaturated soils. Investigations showed that infiltration recharge (mm year^{-1}) is 3–10 at dry fields, 10–32 at settlements, and ranges from –50 to 300 at irrigated fields depending on crops, irrigation technique, inter-vegetative period watering as well as the thickness and lithology of the unsaturated zone (in dry periods negative recharge values occur at a water table depth of less than 2 m in the absence of drainage).

Key words flow model identification; irrigation; real and predicted infiltration recharge; water and salt transport; Ukraine

INTRODUCTION

Infiltration recharge is one of the main components of the groundwater balance reflecting surface water conditions. Quantification of infiltration recharge is required for estimating the impact of irrigation systems on the environment as well as projecting protective measures.

In the southern Ukraine, real and predicted infiltration recharge distribution was estimated within the Dnieper-Molochnaya interfluve and the Steppe Crimea where the largest active and planned irrigation fields (about one million ha) are located. Initially, infiltration recharge was estimated for typical sites distinguished by factors such as meteorology, land use and modification, geomorphology, geology, and hydrogeology, by applying the following direct or indirect measurements and calculations: (a) studying the water balance in the shallow and vadose zones as well as at the surface for several sites or microcosms (Lebedev, 1976; Rode, 1960); (b) comparison of groundwater and infiltrating water concentrations for steady state conditions (Sitnikov, 1986); (c) calculation of water transport velocity in the vadose zone as $V = kI$ where hydraulic conductivity (k) is determined using capillary pressure measurements at the field-scale and a $k(P)$ relationship obtained by laboratory experiments. Measured capillary pressure was used to calculate the moisture gradient $I = (\Delta P - \Delta z)/\Delta z$ where z is the distance between measurement points.

However, since the experiments were constrained by labour and cost limitations, we managed to obtain only several approximate values; these were used as limitation criteria for flow model identification and inverse task solution when determining the areal distribution and extent of current recharge.

CURRENT RECHARGE

Current infiltration recharge was estimated during flow model identification by means of inverse problem solution using stochastic nonlinear programming (Ognianik & Paramonova, 1999). Discrete infiltration recharge values were determined on a $4 \times 4\ \mathrm{km}^2$ grid for an area of 45 000 km^2. Infiltration recharge (mm year^{-1}) is 3–10 at dry fields, 10–32 at settlements, and ranges from –50 to 300 at irrigated fields depending on crops, the irrigation technique, watering during inter-vegetative periods, as well as the thickness and lithology of the unsaturated zone, and presence of drainage.

PREDICTED RECHARGE

Predicted infiltration recharge and its concentrations were studied using water and salt transport models reflecting the lithology and salinization of the unsaturated zone:

$$\begin{cases} \dfrac{\partial\Theta}{\partial H}\dfrac{\partial H}{\partial t} = \dfrac{\partial}{\partial z}\left[k(\Theta)\dfrac{\partial H}{\partial z}\right] \\ \dfrac{\partial(\Theta C)}{\partial t} = \dfrac{\partial}{\partial z}(D^{*}\dfrac{\partial C}{\partial z}) - \dfrac{\partial(vC)}{\partial z} \end{cases}$$

where Θ is volumetric moisture content. $H = P(\Theta) - z$ is a force function, $P(\Theta)$ is capillary pressure, z is a vertical coordinate ($z = 0$ at the surface and is positive downward), $k(\Theta)$ is water hydraulic conductivity, D^* is a dispersion coefficient, C is a pore water solution concentration, and t is time.

$v = -k(\Theta)\ \partial H/\partial z$ gives a water-transport velocity

$W = v\Delta t$ represents an infiltration recharge for Δt period

At the surface, the vegetation and inter-vegetation period conditions were modelled corresponding to an eleven-course rotation with the following irrigation rates ($\mathrm{m}^3\ \mathrm{ha}^{-1}$): 2950 for maize silage, 2750 for winter crops with planted-after maize, 2400 for grain maize, 3150 for spring crops and planted-after-one-year lucerne, 3650 for two-year lucerne, 3400 for three-year lucerne, 3650 for four-year lucerne, 3650 for winter wheat with planted-after-one-year grass, 2600 for tomatoes, 2800 for cabbage, 2900 for forage roots. During irrigation, the watering rate was assigned without evaporation loss, whereas in the inter-watering period a step-by-step moisture decrease was specified in the upper half-metre layer until 0.75 of field water capacity was reached. In the inter-vegetation period a boundary condition is retained where V = constant = $\Delta A/\Delta t$ where ΔA is the difference between precipitation and evaporation during period Δt. A 17-year time series was specified with ΔA varying from 207 mm to –220.1 mm. A condition corresponding to downward flow determined using previous modelling data was assigned at the lower boundary. It allowed groundwater table fluctuations to be modelled. In the case where shallow groundwater was absent, i.e. where loam and clay are underlain with limestone, gravel, or shingle, in which flow is interrupted due to the abrupt increase of water transport velocity, a free gravitational drainage condition was assigned at the lower boundary. $P(\Theta)$ and $K(\Theta)$ relations used for modelling were obtained by laboratory experiments (Golovtchenko *et al.*, 1983). Table 1 demonstrates the maximum, minimum, and average infiltration recharge values depending on lithology of the unsaturated zone and groundwater table depth.

Table 1 Infiltration recharge for different depths and lithology of the unsaturated zone.

Lithology	Water table depth (m)	Recharge (mm year^{-1}):		
		Min	Max	Average
1. Medium loam	7	33	182	79
	5	28	191	74
	2.5	15	231	69
	2	–43	227	55
	1.9	–70	224	50
	1.8	–221	171	11
	1.7	–247	148	–14
2. Medium loam underlain with strongly permeable rocks. BD = 5 m	>5	41	176	81
BD = 3 m	>5	35	169	83
BD = 1 m	>5	38	240	90
3. Medium loam ($BD \leq 3$ m)	20–25	25	120	60
Sandy loam ($BD \leq 8$ m)	20–25	25	120	60
Heavy loam (BD = 20–25 m)	20–25	25	120	60
4. Medium loam ($BD \leq 2.5$ m)	22	22	110	55
Light loam ($BD \leq 5.5$ m)	22	22	110	55
Medium loam ($BD \leq 9.5$ m)	22	22	110	55
Heavy loam ($BD \leq 22$ m)	22	22	110	55
5. Medium loam ($BD \leq 8.5$ m)	18	20	100	50
Heavy loam ($BD \leq 18$ m)	18	20	100	50
6. Heavy loam	>7	15	80	44
7. Clay with a hydraulic conductivity value of 0.002 m day^{-1}	>7	10	60	36
8. Heavy loam (BD = 4–6 m) underlain with sand	>7	25	110	52

BD is the bottom depth.

Figure 1 shows several pore solute concentration diagrams during long-term irrigation for different lithologies and thickness of the unsaturated zone.

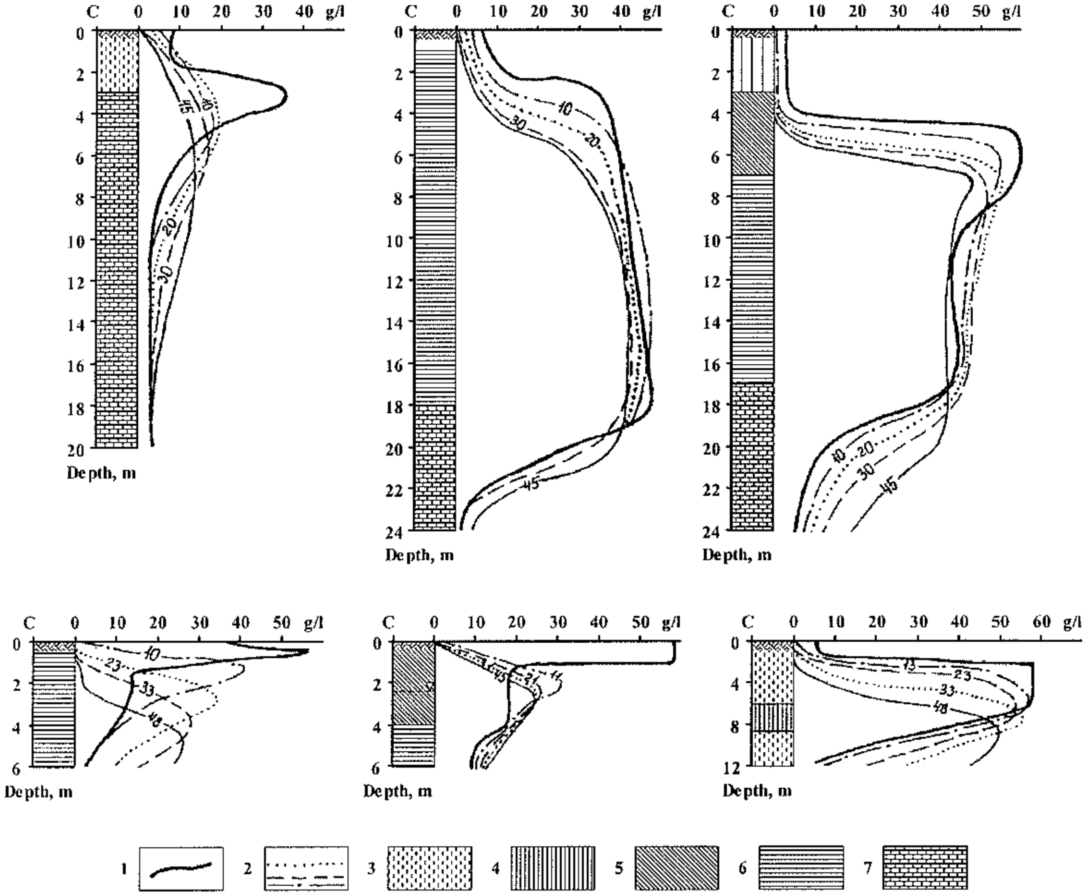

Fig. 1 Pore solute concentration diagrams. 1: real diagrams, 2: predicted diagrams (a number denotes years of irrigation), 3: medium loess loam, 4: heavy loess loam, 5: heavy loam, 6: clay, 7: limestone.

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