

FLOODING AND PLANT GROWTH
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CHAPTER 2. EFFECTS OF FLOODING ON SOILS

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PHILIPPINES

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Flooding or submerging in water an air-dry soil sets in motion a series of physical, chemical, and biological processes that profoundly influence the quality of a soil as a medium for plant growth. The nature, pattern, and extent of the processes depend on the physical and chemical properties of the soil and on duration of submergence. Draining and drying a flooded soil reverses most of those changes.

I. Physical

A. Restriction of gas exchange

1. Gas exchange in well-drained soils

Ten to sixty percent of the volume of soils supporting dry-land plants is gas. The proportion of gas to water and solids is high in free-draining, arable soils with good structure. The composition of gas in a well-drained soil is fairly stable, in spite of oxygen consumption, carbon dioxide production, and nitrogen fixation by soil organisms, because of rapid exchange of gases between soil and air.

Gas exchange is due to mass flow caused by variations in temperature, pressure, wind speed, or rainfall and to diffusion. Gaseous diffusion is by far the more important of the two processes, as cited by Baver et al. (1972) and Russell (1973).

Application of Fick's Law

$$dq = -DA \left(\frac{dc}{dx} \right)_T dt \quad (1)$$

(where D is the diffusion coefficient, dq the number of moles of the substance diffusing in time dt across a cross-sectional area A under a concentration gradient of dc/dx at temperature T) to gaseous diffusion in soils has shown that the gas in a porous, well-drained soil can be maintained at near equilibrium with the atmosphere (Baver et al., 1972). Measurements of oxygen and carbon dioxide fluxes confirm that (Greenwood, 1970).

2. Gas exchange in flooded soils

Increasing the water content of an air-dry soil drastically decreases the term A in Eq. (1). Thus gaseous diffusion virtually ceases when the gaseous pore space drops to about 10% (Wesseling, 1974). That occurs at soil moisture tensions of about 20 to 30 cm. Flooding or waterlogging a soil virtually eliminates gas-filled pores and limits gas exchange between soil and air to molecular diffusion in soil water. This process is about 10^4 times slower than in air, for $D_{\text{water}}/D_{\text{air}}$ for both oxygen and carbon dioxide is 1.13×10^{-4} (Grable, 1966). Consequently the oxygen supply of the soil is cut off and gases formed by soil metabolism accumulate.

3. Depletion of molecular oxygen

Within a few hours of flooding, microorganisms and roots use up the oxygen present in the water or trapped in the soil and render a submerged soil practically devoid of oxygen. I have summarized

the evidence for the absence of molecular oxygen in flooded soils and sediments (Ponnamperuma, 1972).

A flooded soil, however, is not uniformly devoid of oxygen. The oxygen concentration may be high in a surface film or layer, not more than a few mm thick, which is in contact with the oxygenated surface water. The thickness of the layer represents a balance between oxygen diffusion into the soil and its consumption chemically and biochemically. The oxidized layer has the chemical characteristics of an aerobic soil and acts as a sink for substances diffusing upwards from the anaerobic bulk of the soil. The aerobic/anaerobic interface has important implications for the nitrogen economy of flooded soils.

4. Accumulation of soil gases

The drastic restriction of gas exchange between a flooded soil and the atmosphere leads to the accumulation in the soil of nitrogen, carbon dioxide, methane, and hydrogen. The gases build up pressure and escape as bubbles. Analyses of such bubbles from flooded rice fields at different times during a season revealed wide variations in composition, as follows: nitrogen, 10 to 95%; methane, 15 to 75%; carbon dioxide, 1 to 20%; and hydrogen, 0 to 10% (Harrison and Aiyer, 1913). More recent work has shown that carbon dioxide concentrations as high as 50% may persist for several weeks in cold, acid soils that are flooded. (Ponnamperuma, 1976).

B. Thermal

1. Net solar radiation absorbed

Flooding affects net solar radiation absorbed by a soil, heat fluxes in and out of a given volume of soil, the heat capacity of the soil, and soil temperature.

Net solar radiation absorbed by a soil (H) is related to the incident irradiance (R), the albedo of the surface (s), and long-wave radiation (B) according to the following equation:

$$H = R(1-s) - B \quad (2)$$

Because water darkens a soil's color and because water surfaces have a low reflectancy, flooding reduces the value of s in Eq. (2). Some albedo values adapted from Baver et al. (1972) are as follows: dry clay, 0.23; wet clay, 0.16; and water, 0.03-0.10.

2. Heat capacity and heat flux

The specific heat capacity of a soil (C) can be related to the volume fractions of soil minerals (V_1), organic matter (V_2) and water (V_3) and their specific heats by

$$C = 0.46 V_1 + 0.60 V_2 + V_3 \quad (3)$$

and the heat flux density (dQ/dt) by

$$dQ/dt = -kdt/dx \quad (4)$$

Water content affects thermal conductivity k , and consequently the thermal diffusivity k/C depends on water content, that is determined by the separate dependencies of k and C . Usually k/C is lessened by flooding, and both diurnal and seasonal fluctuations of soil temperature are consequently damped.

3. Soil temperature

The interaction of net radiation, absorption, heat fluxes, thermal diffusivity, and the temperature of the incoming flood water determine soil temperature. Figure 1 shows the temperature profile of a flooded rice field (Kondo, 1952). The mean temperature of the surface layer was 6°C lower for flooded soils than for well-drained soils (Bonneau, 1982).

Fig. 1

The temperature of a flooded soil markedly affects: the velocity and pattern of chemical and electrochemical changes initiated by flooding, the release of nutrients, production of toxins, and plant growth (Ponnamperuma, 1976).

C. Swelling of colloids

When a dry soil is flooded, soil colloids absorb water and the soil swells. The rate of water sorption and volume increase of mineral soils depend on the clay content, type of clay mineral, and the nature of the exchangeable cations. Swelling is usually complete in 1 to 3 days. The higher the clay content the greater

the swelling. The expanding-lattice type of clays (smectite and beidellite) swell more than the fixed-lattice type (kaolinite and halloysite). Sodic soils swell more than calcareous or acid soils.

D. Rheological changes

1. Consistency

As the moisture content of a dry soil is increased, cohesion of water films around soil particles causes them to stick together rendering the soil plastic. At this moisture content soils are easily puddled. At higher water contents (as in flooded soils) cohesion decreases rapidly. The rapid decrease is due to increase in the thickness of water films between soil particles as shown by

$$F = \kappa 4 \pi r T/d \quad (5)$$

where F = cohesion, κ = a constant, r = radius of particles, T = surface tension, and d = distance between particles.

Loss of cohesion decreases soil strength rendering the use of heavy machinery on flooded soils impractical.

When a flooded soil is drained and dried, cohesion increases, and the soil shrinks and cracks. Alternate flooding and drying helps aggregate formation.

2. Plasticity

Soil colloids absorb water, swell, and lubricate the coarser particles. That makes a soil moldable or plastic. The plastic

limit is the moisture content at which the soil becomes plastic; the liquid limit signifies a moisture content at which the water films are so thick that the soil mass flows under slight applied pressure (Brown, 1977). Increase in clay content increases the limits. So do smectite, exchangeable sodium, and organic matter.

Flooded soils are well above the liquid limit. So flooded soils have low shear strength or resistance to slipping and sliding of soil particles one over the other and also low compression or change in volume under an applied stress. Subjecting a flooded soil to compression or shearing forces aggravates structure impairment that follows the flooding of a dry soil.

E. Destruction of structure

Good soil structure is vital to dryland crops because it helps aeration and drainage. Flooding a dry soil destroys soil structure by disrupting the aggregates whose arrangement constitutes structure.

The breakdown of aggregates is due to reduction in cohesion with increase in water content, deflocculation of clay due to dilution of the soil solution, pressure of entrapped air, stresses caused by uneven swelling, and destruction of cementing agents. Sodic soils show marked aggregate breakdown on flooding, whereas soils high in iron and aluminum oxides or organic matter suffer little aggregate destruction (Sanchez, 1976). On drying and re-oxidation, structure is regenerated through soil cracking and cementing by hydrated ferric oxides.

F. Water movement

Water moves out of the root zone of flooded soils by evapotranspiration and by percolation and seepage. Evaporation from a free water surface depends on the net radiant energy available at the surface, on the atmospheric temperature and humidity, and on the wind speed; plants growing in an expanse of water have a slight influence on the rate of evaporation of that water.

Percolation is the downward movement of water under the influence of gravity. Seepage is the lateral movement of subsurface water. Because water movement in flooded soils is a combination of the two processes, Wickham and Singh (1978) used seepage and percolation collectively.

Factors that favor water movement in flooded soils under a fixed head of water are a coarse texture, granulated structure, low bulk density, 1:1 type clays, calcium saturation, organic matter, soil cracking, absence of plow pans, and a deep water table (Wickham and Singh, 1978). Flooding decreases water movement in soils of low permeability because of dispersion of soil particles, swelling, aggregate destruction, and clogging of pores by microbial slime. In porous, non-swelling soils, flooding (by providing a greater head of water) increases percolation.

Percolation at more than 1.5 m per day is necessary to meet the oxygen demands of roots in a flooded soil (Grable, 1966). Percolation at over 10 mm per day improves rice yields (Greenland, 1981). This is attributed to removal of toxins. Percolation at 2-3 cm per day does not oxygenate more than 1 cm of the surface layer of a flooded soil (Houng, 1981).

II. Biotic

The presence of a layer of standing water and the differentiation of the soil profile into aerobic and anaerobic zones have profound consequences on the ecology of flooded soils.

A. Standing water

The surface water layer becomes the habitat of heterotrophic flora and fauna, as well as algae and macrophytes. Consequently it is a zone of intense physiological activity.

The bacteria are aerobic and the fauna largely zooplankton (Watanabe and Furusaka, 1980). The algae include phytoplankton, filamentous algae, and higher algae growing as anchored species (Roger and Watanabe, 1983). Physiologically, they are either nitrogen fixers or non-nitrogen-fixers. The macrophytes are of three kinds: surface plants both rooted and free-floating; submerged rooted plants growing largely below the surface; and plants growing in shallow water or wet soil. The biomass production in the standing water of fallow tropical rice fields may be as high as 12 t fresh matter per ha per 100 days (Roger and Watanabe, 1983).

The main physiological activities in the standing water are photosynthesis, respiration, ammonification, nitrification, and nitrogen fixation. Because of photosynthesis by algae and macrophytes during daytime, the water gets supersaturated with oxygen and unsaturated with carbon dioxide. Depletion of carbon dioxide may increase the pH to 10.0 (De Datta, 1981) during daytime. At night the oxygen concentration decreases because of

oxygen consumption by respiration. Cyanobacteria (blue-green algae) may fix as much as 0.5 kg N/ha/day (Stewart et al., 1979). Ammonia formed by mineralization of organic matter or diffusing upwards from the anaerobic zone is nitrified.

B. Aerobic/anaerobic interfaces

A flooded soil has three types of aerobic-anaerobic interfaces: the surface oxidized layer in contact with oxygenated water and the anaerobic soil matrix; the rhizosphere of marsh plants; and the oxidized subsoil.

The surface oxidized layer receives plant debris and mineral matter from the supernatant water and substances diffusing upwards from the anaerobic zone. That leads to enrichment of 1 cm of the surface soil with organic matter, nitrogen, phosphorus, iron, manganese, and other substances. Aerobic bacteria, algae, and photosynthetic bacteria are present. In addition, methane oxidizing bacteria oxidize methane moving into the surface layer from the anaerobic layer (Harrison and Aiyer, 1915; De Bout et al., 1978).

Because of oxygen diffusion from the roots of marsh plants, the rhizosphere is oxidized (Armstrong, 1978). Thus the rhizosphere of rice has a positive effect on aerobic bacteria including heterotrophic nitrogen fixers and a negative effect on Clostridia, sulfate reducers, and denitrifiers (Watanabe and Furusaka, 1980). Yoshida (1978) cited evidence for nitrogen fixation in the rhizosphere of other aquatic plants and microbial sulfide oxidation

in rice. Diffusion of organic substrates and nitrogen from rice roots helps nitrogen fixation in flooded rice fields (Ponnamperuma, 1972; Yoshida, 1978; Buresh et al., 1980).

The formation of manganic oxide deposits below the plow sole of flooded rice fields has been attributed to the presence of manganese oxidizing bacteria (Watanabe and Furusaka, 1980).

C. The anaerobic zone

Within a few hours of flooding an air-dry soil, the bulk of a soil is rendered practically devoid of molecular oxygen. Consequently aerobes are replaced by facultative anaerobes, which in turn are superseded by strict anaerobes (Yoshida, 1978). Fungi and actinomycetes are suppressed and bacteria predominate. Strict anaerobes in a paddy field constituted only 10% of the count of facultative anaerobes (Watanabe and Furusaka, 1980). Soil metabolism which causes denitrification, reduction of manganese, iron, and sulfate, as well as nitrogen fixation and methane formation is the work of anaerobic bacteria in the oxygen-free zone of flooded soils (Yoshida, 1978; Watanabe, 1983).

III. Electrochemical

Flooding air-dry soils causes direct and indirect changes. One direct and almost instantaneous change is dilution of the soil solution. That increases pH, decreases electrical conductance, and alters the diffuse double layer of colloidal particles. But these changes are insignificant compared with the drastic changes in redox potential, pH, electrical conductance, ionic strength, ion exchange, and sorption and desorption caused by soil reduction.

Flooding cuts off a soil's oxygen supply. After the aerobic organisms have used up the oxygen present in the soil, facultative and obligate anaerobes proliferate using oxidized soil components and dissimilation products of organic matter as electron acceptors in their respiration reducing a soil in thermodynamic sequence (Table 1). The reduced state of a flooded soil is shown by its low redox potential (0.2 to -0.4V) compared with 0.8 to 0.3V for aerobic soils, by the absence of NO_3^- , and the presence of Fe^{2+} , Mn^{2+} , NH_4^+ , and S^{2-} .

Table 1

A. Decrease in redox potential (Eh)

1. Eh changes

Redox potential falls rapidly after flooding, reaches a minimum within a few days, rises rapidly to a maximum and then decreases asymptotically with time (Fig. 2 and 3). The presence of organic matter and a temperature of 35°C favor Eh decrease and values as low as -0.25V may be reached in soils low in nitrate and in manganic and ferric oxides, within 2 weeks after submergence (Ponnamperuma, 1972; 1981).

Fig. 2, 3

Gambrell and Patrick (1978) have given the following potentials at pH 7.0 for the observations listed below:

Observation	Eh (V)
Disappearance of O_2	0.33
Disappearance of NO_3^-	0.22
Appearance of Mn^{2+}	0.20
Appearance of Fe^{2+}	0.12
Disappearance of SO_4^{2-}	-0.15
Appearance of CH_4	-0.25

Eh is high in the flood water and in the first few mm of the surface soil. Then it drops rapidly with depth to negative values. Positive values are rapidly resumed below the plow sole where Mn(IV) compounds are present (De Gee, 1950).

The Eh is a useful guide to the sequence of soil reduction, but it does not indicate the capacity of the soil to resist Eh changes. Buffering agents in the oxidizing range are oxygen and nitrate, whereas in the reducing range ferric compounds dominate (Patrick, 1981).

2. Theoretical

Although redox potentials of flooded soils have been used in ecological studies (Pearsall and Mortimer, 1939; Bass-Becking et al., 1960; Aomine, 1962; Ponnampereuma, 1981), they are of limited theoretical or practical value. But potentials of interstitial waters of flooded soils and sediments are thermodynamically meaningful and have been used successfully to explain the changes in concentration of ecologically important ions in soil solutions (Ponnampereuma, 1972; Yamane, 1978; Lindsay, 1979; Rowell, 1981).

For the equilibrium



$$Eh = Eo + 2.303 \frac{RT}{nF} \frac{(\text{Ox})}{(\text{Red})} - 2.303 \frac{RTm}{nF} \text{pH}$$

$$\text{or } \text{pE} = \text{pEo} - \frac{1}{n} \text{p}(\text{Ox}) + \frac{1}{n} \text{p}(\text{Red}) - \frac{m}{n} \text{pH} \quad (7)$$

where (Ox) and (Red) are the activities of the oxidized and reduced phases, Eh is the redox potential, pE (electron activity) = Eh / 2.303 RTF⁻¹ = Eh / 0.0591 at 25°C, and the other symbols have the usual significance.

The solutions of flooded soils have Eh values of -60 to 170 mV or pE values of -1 to +3 at pH 7. The Eh or pE values of the solutions of most acid, mineral soils that have undergone reduction conform to the equations

$$\text{Eh} = 1.06 - 0.059 \log \text{Fe}^{2+} - 0.177 \text{ pH} \quad (8)$$

$$\text{or} \quad \text{pE} = 17.87 + \text{pFe}^{2+} - 3 \text{ pH} \quad (9)$$

B. Changes in pH

1. The pH values of flooded soils

When an acid soil is kept flooded its pH increases whereas the opposite happens in alkaline soils (Fig. 4). Thus the pH values of most flooded mineral soils lie between 6.7 and 7.2; those of their interstitial solutions are between 6.5 and 7.0 (Ponnamperuma, 1972). But if a soil's reducible iron content is low, its pH may not rise above 5.0 even after months of submergence, and if the organic matter content of a high pH soils is low, the pH may not decrease below 8.0.

Fig. 4

The increase in pH of acid soils is due mainly to the reduction of Fe(III) to Fe(II). It can be described for most acid, mineral soils by Eq. 7 or 8. The decrease in pH of sodic and calcareous soils and the check on the pH rise of acid soils are due to the accumulation of carbon dioxide. Some observed relationships between pH and the partial pressure of carbon dioxide (P_{CO_2}) are as follows (Ponnamperuma et al., 1969):

$$\text{Sodic soil} \quad \text{pH} = 6.4 - 1.00 \log P_{\text{CO}_2} \quad (10)$$

$$\text{Calcareous soil} \quad \text{pH} = 6.1 - 0.66 \log P_{\text{CO}_2} \quad (11)$$

$$\text{Neutral soil} \quad \text{pH} = 6.1 - 0.64 \log P_{\text{CO}_2} \quad (12)$$

$$\text{Ferruginous acid soil} \quad \text{pH} = 6.1 - 0.57 \log P_{\text{CO}_2} \quad (13)$$

According to Eqs. (10-13) at a P_{CO_2} of 0.1 bar (a common value in flooded soils) the pH of the soils is 6.7 to 7.4

2. Effects of pH on chemical equilibria in flooded soils

A soil's pH markedly affects the concentration in the soil solution of ecologically important ions through the involvement of hydrogen ions in chemical equilibria (Ponnamperuma, 1972; 1978; 1981).

The effect of pH on hydroxide equilibria can be described by the following equation:

$$\text{pH} + 1/2 \log M^{2+} = K \quad (14)$$

where M is Fe, AlOH, Zn or Cu, and K = 5.4 for Fe, 2.2 for AlOH, 3.0 for Zn and 1.6 for Cu.

According to Eq. (14) a pH increase of 1 unit should decrease the concentration of M 100 times. Thus an increase of pH from 4.0 to 5.0 depresses the concentration of water-soluble aluminum from 7 to 0.07 mg/L and abolishes aluminum toxicity. On the other hand a pH decrease from 7.0 to 6.5 in a ferruginous soil may increase the concentration of water-soluble iron from 35 to 350 mg/L and cause iron toxicity.

The effect of pH on the solubility of calcium and manganese is given by

$$\text{pH} + 1/2 \log M^{2+} + 1/2 \log P_{\text{CO}_2} = K \quad (15)$$

where M = Ca or Mn, and K = 4.92 for Ca and 4.06 for Mn. At a high pH (8.5) and low P_{CO_2} (0.05 bar) the concentrations of calcium and manganese are 0.08 and 0.003 mg/L respectively. These are inadequate for plant growth. But if as a result of flooding pH decreases to 7.0, the concentrations reach the adequate values of 80 and 3 mg/L respectively.

C. Changes in specific conductance and ionic strength

1. Specific conductance

The specific conductance (κ) of an aqueous solution at constant temperature depends on the nature and content of ions present in solution. In flooded soils κ represents a balance between ion production during soil metabolism and their inactivation or replacement by slower moving ions.

The specific conductance of the interstitial solutions of flooded soils increases during the first few weeks after flooding, reaches a peak, and then declines to a fairly stable value but its kinetics varies with the soil (Fig. 5). The peak increase is 1-2 dS/m.

Fig. 5

The increase in κ during the first few weeks of submergence is due to the production of NH_4^+ , HCO_3^- , RCOO^- , Mn^{2+} and Fe^{2+} followed by the displacement of Na^+ , K^+ , Ca^{2+} and Mg^{2+} from the soil colloids by Mn^{2+} and Fe^{2+} . The decrease is due to removal of

HCO_3^- , conversion of RCOO^- and HCO_3^- to CH_4 , and precipitation of Mn^{2+} as MnCO_3 and Fe^{2+} as $\text{Fe}_3(\text{OH})_8$. The close similarity between the kinetics of alkalinity, cations, and κ show the role of the ions listed above in controlling the changes in κ (Fig. 6).

Fig. 6

2. Ionic strength

The ionic strength (I) of an aqueous solution is related to the concentration (c) and valence (z) of the ions present by

$$I = 1/2 \sum c_i z_i^2 \quad (16)$$

Flooding increases the concentration of NH_4^+ and HCO_3^- . It also converts insoluble Mn(IV) and Fe(III) compounds to Mn(II) and Fe(II) forms which are much more soluble in water. These processes increase ionic strength.

Because the kind and concentration of ions influence both specific conductance and ionic strength, κ and I should be related. For the solutions of reduced soils I (mol/L) is numerically equal to 16κ (dS/m) up to ionic strengths of 0.05 (Ponnamperuma et al., 1966; Yamane, 1978).

D. Ion exchange

Ion exchange is the replacement by ions in the soil solution of cations and anions held by electrostatic attraction on soil colloids.

Soil reduction that follows flooding affects ion exchange reactions by altering the electric charge on soil colloids and by increasing markedly the concentrations in the soil solution of Fe^{2+} and Mn^{2+} in acid soils, Ca^{2+} in alkaline soils, and HCO_3^- in all soils.

1. Cation exchange

Cations are held on soil colloids by permanent negative charges residing on the particles and by pH-dependent negative charges. The permanent negative charge on a mineral colloid is constant in the pH range 2.5 to 10 (Talibudeen, 1981). But the pH-dependent negative charges, arising from the ionization of carboxyl and phenolic groups of soil organic matter and the dissociation of surface hydroxyls in silica and silicates increase as pH increases.

The increase in pH of acid soils on flooding from <3 to about 7 should increase the ionization of carboxyl groups (whose pK_a values are generally between 3 and 5) and of hydroxyls attached to silicon (whose pK_a values range from 2 to 9.7 depending on the mineral). The decrease in pH of alkaline soils should depress the ionization of both groups of colloids. Consequently, flooding should increase the cation exchange capacity of acid soils and decrease that of alkaline soils.

The Fe^{2+} , Mn^{2+} , and NH_4^+ concentrations in the solutions of flooded soils may attain values as high as 10, 2, and 8 mM, respectively. These ions displace Na^+ , K^+ , Ca^{2+} , and Mg^{2+} into the soil solution. The close parallelism between the kinetics of water-soluble Fe^{2+} and Mn^{2+} and other cations illustrates this (Fig. 6). The findings of Sims and Patrick (1978) that greater amounts of exchangeable iron, manganese, zinc, and copper were present at low Eh and low pH than at high Eh and pH confirm the role of Fe^{2+} and Mn^{2+} in cation exchange in flooded soils. Low Eh and low pH favor the mobilization of Fe^{2+} and Mn^{2+} .

2. Anion exchange

At pH values below their point zero charge, hydrous oxides of aluminum, iron, and manganese are positively charged. Below pH 6, exposed oxygen and hydroxyl groups in 1:1 and 2:1 lattice clays accept protons and acquire a positive charge. Amino groups of soil organic matter carry positive charges below their pK_b values (Talibudeen, 1981; Mott, 1981).

The positive sites attract anions. Of common anions, NO_3^- and Cl^- are held by simple electrostatic attraction and are said to be nonspecifically adsorbed. All other anions including organic anions, are specifically adsorbed and do not readily undergo ion exchange (Mott, 1981).

Flooding affects the anion exchange properties of soils by: increasing the pH of acid soils and decreasing that of alkaline soils; reducing Fe(III) and Mn(IV) hydrous oxides to the Fe(II) and Mn(II) forms; converting NO_3^- to N_2 and SO_4^{2-} to insoluble sulfides; and loading the soil solution with up to 50 mM HCO_3^- .

The increase in concentration of water-soluble phosphate and silica and desorption of sulfate caused by flooding soils (Ponnamperuma, 1972) may be due to decrease in anion exchange capacity and increase in the bicarbonate concentration.

E. Sorption and desorption

Sorption is the concentration of gases, liquids, or solutes in a layer of molecular dimensions on surfaces of solids with which they are in contact.

Sorption is due to cation or anion exchange, covalent bonding, action of van der Waals forces, or isomorphous substitution. When the mechanism is known, the process is called adsorption. Sorbed substances are in equilibrium with the soil solution and may undergo desorption. The sorbing agents are clay, hydrous oxides of aluminum, iron, manganese, and silicon, or organic matter (Ellis and Knezek, 1972; Stevenson and Ardakani, 1972).

Sorbates of ecological interest include: the cationic forms of copper, zinc, and metallic pollutants; the anionic forms of boron, molybdenum, phosphorus, and sulfur; and organic substances, including pesticides. The Freundlich and Langmuir isotherms are used to describe sorption. According to both equations sorption increases with concentration of the substance sorbed.

Flooding dilutes the soil solution. It should therefore favor desorption. But the reduction of Fe(III) and Mn(IV) hydrous oxides drastically alters their surface properties and produce large amounts of water-soluble Fe^{2+} and Mn^{2+} . Thus ions sorbed on them may be desorbed and also exchangeable cations may be displaced into the solution. The increase in pH of acid soils and the decrease in pH of alkaline soils affect the surface properties of clays, hydrous oxides, and organic matter. That may increase sorption or desorption, depending on the point zero charge of the sorbent.

Sorption and desorption play an important role in controlling the availability of plant nutrients and pollution by toxic metals and organic pesticides (Burchill et al., 1981).