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# FACTORS AFFECTING REDOX REACTIONS IN HYDROMORPHIC SOILS. A REVIEW

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Hydromorphic soils are found around the globe and they are well distinguished from other soil classes. The nature and severity of hydromorphism in these soils is determined by redox reactions that are highly dependant on the level and duration of soil saturation with water. Redox potentials are good indicator of redox reactions and/or hydromorphism. However, these redox reactions and hydromorphic conditions are directly affected by the presence or absence of various factors such as oxygen, organic matter and micro-organisms. In this manuscript a comprehensive review is presented about the hydromorphic soils, causes of hydromorphism, measurement of hydromorpic conditions, factors affecting the hydromorphic conditions and how the redox reactions can play an important role in pedogenesis of hydromorphic soils.

**Keywords:** Hydromorphic soils, redox reactions, soil compaction, waterlogging, hydromorphism

## INTRODUCTION

Generally speaking, soil is roughly consisted of 50 % solids, 25% water and 25 % air. Furthermore, basic composition of soil air consists of nitrogen (78%), oxygen (18-20%), carbon dioxide (1-10%), vaporized water (concentration is temperature dependant) and trace gases. However, soil air can be replaced by water when soil is permanently or temporarily saturated with water and if this saturation period is long enough to eliminate the oxygen from the soil, reducing conditions can occur in that soil. This state of permanent or temporary water saturation in the soil associated with conditions of reduction is considered as hydromorphism and this state has considerable effects in the composition, properties, genesis and evolution of the soil (Jacob and Otte, 2003). Various terms have been used in considerable systems of soil description and soil classification such as : hydromorphic properties, mottled horizons, flecks with chroma lower than 2, decoloration, aquic moisture regimes, gleyification, pseudo- gleyification, redoximorphic features (Ponnamperuma, 1972; Jacob and Otte, 2003).

Hydromorphic soils are defined by Soil Science Society of America in the glossary of soil science as "A suborder of intrazonal soils, consisting of seven great soil groups, all formed under conditions of poor drainage in marshes, swamps, seepage areas, or flats". These soils are characterised by the phenomena of reduction and segregation of iron, linked to temporary or permanent saturation of pores by water causing a prolonged deficiency of oxygen (Duchaufour, 1997). According to WSR (2001), hydromorphic soils are classified under the soil reference group of Gleysols (from Russian local name *gley*, mucky soil mass) and occupy an area of 720 million hectares world-wide. Common international names are 'Gleyzems' and 'meadow soil' (Russia), Gleissolos (Brazil), 'Aqu-' suborders of Entisols, Inceptisols and Mollisols (USA), 'Gley' (Germany), and 'groundwater soils' and 'hydromorphic soils'. Many of the WRB (World Reference Base for Soil Resources) Gleysols correlate it with the aquic suborders of the US Soil Taxonomy (*Aqualfs*, *Aquents*, *Aquepts*, *Aquolls*, etc) (WSR, 2006).

Hydromorphic soils and processes affecting the redox reactions are extensively studied and published but there are very few articles which provide comprehensive information and findings about the factors affecting oxido-reduction processes in hydromorphic soils. The objective of this review is to discuss the causes of hydromorphism and various factors which affect the redox reactions in the hydromorphic soils.

***Causes of hydromorphism:*** Hydromorphism can be a result of natural conditions or anthropic influences. It can be direct result of precipitations or an indirect result of heavy precipitations in the forms of surface flooding or water-table rise. In the deltaic and coastal regions, soils are directly influenced by high water tables of fresh water (river) or salty water (sea water). The examples of natural hydromorphism are natural wetlands and humid zones situated at lower altitudes and/or near to a river or a sea. According to Convention of Ramsar (1971) the humid zones are protected and defined as: "The vast expanses of marshes, peat bogs or natural or artificial waters, permanent or temporary where the water may be stagnant or running, brackish or salty; also included are the expanses of marine water where the height of low tide does not exceed six meters". The natural humid zones can be influenced by nature origin phenomenon like subsidence, climatic change, sea level rise, storms and droughts or by human activity like cultivation, roads and dams construction and pumping of ground water (Bijlsma, 1996).

Hydromorphism can be induced by human activity either by excessive irrigation on poorly drained soils or by soil compaction (Herbauts *et al.*, 1996; Qureshi *et al.,* 2008). When a soil with an impermeable horizon at some depth receives the irrigation water (and/or seepage from canals) without proper drainage practices, eventually, the water table rises slowly. That rise in the water table can't be observed in early stages until it is very late and soil is waterlogged. The recent example of this type of waterlogging can be observed in Pakistan, where canalisation of Indus delta in British era (after 1857) not only resulted in waterlogging but also in soil salinity due to evaporation of irrigated water (Bokhari, 1980; Qureshi *et al.,* 2008). Now, According to the Agricultural Statistics of Pakistan (2004), the total area under waterlogged soils is 4.85 million hectares with water table depth ranging from 0-3 m below the soil surface where oxido-reductomorphic conditions may exist but no sufficient data are available (Nawaz *et al.,* 1992; Khan, 2010). As in Pakistan, deltaic soils of the most other regions such as Mekong, Nile, Senegal, and Rhone etc. are affected by hydromorphism and salinity that can be seasonal or permanent. Rising of ground water through capillary action and evaporation has resulted in primary salinity of these soils (Murtaza *et al.*, 2011). One of the solutions to combat this problem is the practice of flood irrigation like in rice culture. So, salts are dissolved by irrigated water and evaporation of ground waters is limited. But this can result in the secondary salinity caused by evaporation of irrigated water in confined conditions and by soluble salt mobilisation to cultivated horizons.

Hydromorphic conditions can also be developed due to soil compaction (Herbauts *et al.*, 1996). Soil compaction can be expressed as a process that involves a rearrangement and bringing of solid particles of the soil closer together and consequently an increase in the bulk density of soil (Panayiotopoulos *et al.*, 1994). Some times, the practice of soil compaction is carried out voluntarily in sandy soils either to reduce the water infiltration and/or to provide a better support for plants (Bouwman and Arts, 2000). Lowland rice cultivation is carried out in flooded and voluntarily compacted soil. A hardpan is prepared at certain depth (20-25 cm) of the soil to reduce the infiltration rate of water through the soil to economise the water but a shallow hardpan (less than 10 cm), (Hemmat and Taki, 2003). However, in major cultivated areas and forest soils, soil compaction is often non-voluntary due to mechanical operations (Nawaz *et al.*, 2013). No doubt, voluntary and/or non voluntary soil compaction can result in the increase in the soil bulk density, low permeability, low infiltration rate and ultimately hydromorphic conditions in the soils (Herbauts *et al.*, 1996; Nawaz *et al.*, 2013).

***Redox potential and hydromorphism:*** In hydromorphic soils, depending on the degree and period of water saturation, cycles of aerobic and anaerobic conditions can be observed resulting in the oxidising and reducing conditions. These cyclic changes that induce the redox reactions in the soil result in different redox potential values. The redox potential (Eh) in a soil solution at given point and time is a good indication of the redox status of the soil (Fiedler *et al.*, 2007). The term "redox" is the representative of processes of reduction and oxidation. As reduction is the reception of electron while oxidation is the donation of an electron, a complete redox reaction consists of both types of reactions also called half-reactions (Glinski and Stepniewski, 1985). An example of a redox reaction is given below, in which, $Fe^{3+}$ is reduced to $Fe^{2+}$:

$$\mathrm{FeOOH} + e^- + 3\,H^+ \rightarrow Fe^{2+} + 2\,H_2\,O$$

In the above given example iron in the form of $Fe^{3+}$ is solid while $Fe^{2+}$ is in dissolved form in soil solution, as solid minerals can dissolve and dissolved ions can become mobile when changes in the valence of such atoms occur (Vepraskas and Faulkner, 2001).

According to Schüring *et al.* (2000) for a following redox reaction:

$$\mathrm{Ox} + \frac{n}{2}\,H_{2(g)} = \mathrm{Red} + nH^+$$

in which an oxidant (Ox) is reduced (Red) while $H_{2(g)}$ is oxidised to $H^+$, with the constant:

$$K = \frac{\{Red\}\{H^+\}^n}{\{Ox\}(p_{H_2})^{\frac{n}{2}}}$$

The redox potential under normal condition is:

$$E^\circ{}_H = \frac{(ln10)RT}{nF} \cdot logK$$

where

$E^\circ_H$ = redox potential [V] (in relation to a normal hydrogen electrode) under normal conditions (all activities = 1, $P_{H2}$ = 1 atm, $H^+$ = 1 M).

F = 1 faraday (= 96485 $C.mol^{-1}$)

n = number of exchanged electrons

R = gas constant (= 8.314472 $J.mol^{-1}.K^{-1}$)

T = temperature in K

(ln10) RT/F = 0.059 V (at 25 °C)

Under other conditions the redox potential is:

$$E_H = E^\circ{}_H + \frac{(ln10)RT}{nF} log\frac{\{Ox\}}{\{Red\}}$$

where

$E_H$ = redox potential [V] (in relation to a normal hydrogen electrode).

Above equation is also called NERNST equation that gives the relationship between the redox potential and the activities of the oxidised and reduced species at equilibrium.

Redox potential (Eh) can be measured practically but its measurement requires that equilibrium is established at the electrode and among various redox couples. In soils, the Eh of the soil solution should be stable across the horizon for a certain time period and all electron acceptors should react at a similar rate. But, above mentioned conditions are often not fulfilled in natural waters and soil solution, either due to slow kinetics of the important potential determining redox couples at the electrode or lack of equilibrium among various redox couples. In soil solution Eh can be more unstable as compared to natural water due to fluctuating degree of saturation of soil and heterogeneity in the distribution of organic matter (Schüring *et al.*, 2000; Vepraskas and Faulkner, 2001).

The use of notion "pe" is more convenient to describe the redox state. It is defined as negative common logarithm of the free electron activity (Schüring *et al.*, 2000):

$$\mathrm{pe} = -\log\{e^-\}$$

Relation between redox potential (Eh) and redox intensity (pe) is described as:

$$\mathrm{Eh} = \frac{(ln10)RT}{F}pe$$

or, equivalently:

$$\mathrm{pe} = \frac{Eh}{(ln10)RT/F}$$

Large values of "pe" indicate the presence of oxidised environment and existence of oxidised species in the solution while small or negative values of "pe" indicate reduced conditions (Glinski and Stepniewski, 1985). The redox potential range in natural waters is limited in the negative range by the reduction of $H_2O$ to $H_{2(g)}$ and in the positive range by the oxidation of $H_2O$ to $O_{2(g)}$. So, at pH 7 and atmospheric pressure, the redox potential range is from about -400 to +800 mV that means "pe" ranges from about 14 to -7 (Schüring *et al.*, 2000).

***Heterogeneity in redox conditions in the soil:*** The redox conditions in soils vary widely over short distances because of heterogeneity in soil physical properties, organic matter contents, temperature and degree and period of water saturation (Sparks, 2003). All above mentioned factors result in variation in $O_2$ diffusion, and microbial activity and respiration in soil. This heterogeneity can result in anaerobic microsites with oxidative soil matrix (Glinski and Stepniewski, 1985). However, the redox conditions are relatively more homogeneous in saturated soils and flooded soils than in drier soils (Bohn *et al.*, 1985). Redox processes in the rhizosphere also play an important role in the heterogeneity of redox conditions of soil. Some plant roots are capable of converting the anaerobic environment to aerobic environment in rhizosphere in flooded and saturated soils that is, sometimes, even remarkable by naked eye due to soil colour change owed to oxidation of $Fe^{2+}$ (Begg *et al.*, 1994; Doyle & Otte, 1997; Jacob & Otte, 2003). It is reported that variation in redox potential across a saturated soil horizon can be as much as 600 mV just after the saturation while this variation can be less about 100 mV after two months of saturation. So, at least five replications of redox potential measurements at each point are recommended to assess above mentioned variations (Vepraskas and Faulkner, 2001).

In general, redox potential in soil ranges from approximately +1 to -1 V, being relatively small, it is expressed in mV (Vepraskas and Faulkner, 2001). When soil pores are not filled with water (unsaturated) and permit the free diffusion of $O_2$, oxidised conditions are observed in the soil with relatively high Eh values (more than 500 mV at soil pH 7). When soil becomes water saturated, due to insufficient diffusion of $O_2$, dissolved $O_2$ can be exhausted in 24 hours which results in anaerobic conditions in the soil (McBride, 1994). Depending on the soil pH, in the absence of oxygen, biological reduction of $NO_3^-$ and other elements is carried out. Thermodynamic conditions favour the reduction or oxidation of these elements at different redox potentials. For example at pH 7, thermodynamically, reduction of $SO_4^{2-}$ is favoured at lower redox potentials (less than 0 mV) than $Fe^{3+}$ (less than 300 mV) (Schüring *et al.*, 2000). Reducing reactions for major elements and their concerning redox potentials in the field are presented in table 1.

**Table 1. Reducing reactions for major elements with their accompanying redox potentials in the field (Bohn *et al.*, 1985)**

| Half reaction | Redox potential measured in soil (mV) |
|---|---|
| $O_2 + 4\,e^- + 4\,H^+ \rightarrow 2\,H_2O$ | 600 to 400 |
| $NO_3^- + 2\,e^- + 2\,H^+ \rightarrow NO_2^- + H_2\,O$ | 500 to 200 |
| $MnO_2 + 2\,e^- + 4\,H^+ \rightarrow Mn^{2+} + 2\,H_2\,O$ | 400 to 200 |
| $FeOOH + e^- + 3\,H^+ \rightarrow Fe^{2+} + 2\,H_2\,O$ | 300 to 100 |
| $SO_4^{2-} + 6\,e^- + 9\,H^+ \rightarrow H_2\,S + 4\,H_2\,O$ | 0 to -150 |
| $2\,H^+ + 2\,e^- \rightarrow H_2$ | -150 to -220 |
| $2\,CH_2\,O \rightarrow CO_2 + CH_4$ | -150 to -220 |

***Factors affecting the hydromorphism:*** The most important factor affecting the redox conditions of soil is the presence of oxygen (aerobic environment) or absence of oxygen (anaerobic environment) while other factors such as organic matter, activity and type of microorganisms, temperature, and soil composition can not be neglected. Effect of all these factors on the redox conditions of soil are discussed in detail.

***Soil oxygenation:*** Soil aeration is very important for chemical, physical and biological properties of soil, which ultimately depends on the soil porosity (Cannell, 1977; Czy, 2004). It permits the free exchange of gases, mostly by diffusion, and helps to maintain the balance among different gases in the soil. In the case, when soil porosity is not sufficient (soil compaction) or soil pores are filled with water, the restricted gaseous exchange results in lower diffusion of oxygen than consumption during respiration (Drew, 1990). If this condition persists for a long time as in saturated or flooded soils, after a certain time, $O_2$ disappears from the soil air that can further lead to increase in the concentrations of trace gases ($CH_4$, $H_2S$, $N_2O$, $C_2H_4$, and $H_2$) in the soil (Glinski and Stepniewski, 1985).

Soil porosity, differentiated as air-filled porosity and water-filled porosity, is the best physical parameter to determine the ability of soil to exchange of gases with atmosphere. A decrease in air filled porosity leads to a decrease in oxygen concentration in the soil. An air-filled porosity ($E_g$) is the difference of total soil porosity ($f$) and its current volumetric water contents ($\theta_v$) (Glinski and Stepniewski, 1985):

$$E_g = f - \theta_v$$

the total porosity is calculated as:

$$f = 1 - \rho_b / \rho_s$$

where: $\rho_b$ = bulk density (Mg $m^{-3}$)
$\rho_s$ = particle density (Mg $m^{-3}$)

while volumetric water content ($\theta_v$) can be calculated as:

$$\theta_v = \theta_w . \rho_b / \rho_w$$

where: $\theta_w$ = gravimetric water content (water content by weight), $\rho_w$ = water density (Mg $m^{-3}$), $\theta_v$, $\theta_w$, $f$ and $E_g$ are dimensionless.

The exchange of gases between soil and atmosphere, particularly for oxygen and carbon dioxide, is carried out by pressure gradients (mass flow) and concentration gradients (diffusion flow) (Hillel, 1980). Both kinds of flow may take place in the soil pores but relative contribution of the both may vary within broad limits depending on the atmospheric and soil physical and biological conditions (Glinski and Stepniewski, 1985; Ouyang and Boersma, 1992). Generally, most of oxygen in the soil is exchanged with atmosphere through diffusion flow (Grable and Siemer, 1968). Soil $O_2$ levels are generally more in upper surface of soil as compared to lower soil horizons due to slow diffusion rates of oxygen from the surface through the soil. Furthermore, soils with higher permeability result in more rapid oxygen diffusion into and through the soil, and similarly is the case for carbon dioxide movement (Drew, 1990). A good correlation exists between redox potentials and presence or absence of oxygen (Grable and Siemer, 1968).

Oxygen is the most important component that affects the biological and chemical transformations in the soil (Silver *et al.*, 1999). It influences the microbial population and their mode of respiration (Kowalchuk *et al.*, 1998), enzyme activity, and redox potentials in the soil (Brzezinska *et al.*,1998).

***Organic matter:*** Soil organic matter (SOM) can be defined as "a mixture of plant and animal residues in different stages of decomposition, substances synthesized microbiologically and/or chemically from the breakdown products, and the bodies of live and dead microorganisms and their decomposing remains" and humic substances (HS) are "a series of relatively high-molecular-weight, brown to black substances formed by secondary synthesis reactions". Natural organic matter (NOM) is broken down organic matter that comes from plants and animals in the environment. NOM is assigned to the realm of all of this broken down organic matter. SOM ranges from 0.5% on weight basis in surface horizons of mineral soil to 100% in organic soils (Histosols) (Sparks, 2003). Major components of organic matter include cellulose, hemicellulose, lignin, simple saccharides, amino-acids and aliphatic acids (Glinski and Stepniewski, 1985). Quantity of SOM influences the redox conditions of the soil (Dobos *et al.*, 1990).

In aerobic environment, organic matter can be completely mineralized to carbon dioxide by microorganisms which use oxygen as "Terminal Electron Acceptor" (TEA) but in anaerobic decomposition of organic matter is relatively slow and involves many interdependent and interrelated microbial processes that can produce both carbon dioxide as well as methane (Megonigal *et al.*, 2004). Therefore, the aerobic decomposition is different from anaerobic decomposition of organic matter. In anaerobic decomposition, hydrolysis and fermentation reactions degrade the organic molecules and result in successively lower molecular weight products. The end products of these reactions include acetate, hydrogen and carbon dioxide, which are subsequently used as substrates for microbial respiration. In the absence of oxygen, microbes preferentially reduce a variety of alternative TEAs for respirations, in the order of decreasing thermodynamic yield, such as $NO_3^-$, $Mn^{4+}$, $Fe^{3+}$, $SO_4^{2-}$, and $C^{4+}$ (Keller *et al.*, 2009). When ferric iron is available, ferric iron reduction is an important pathway for organic matter decomposition (Lovley, 1987).

Humic substances, being more reactive, form the organo-inorgano complexes in the soils that are relatively stable and less susceptible to microbial attack (Sparks, 2003). Recently, it is reported that solid phase humic substance can serve as electron acceptors while humic acid can serve as either electron acceptor or electron donor during the decomposition in the absence of oxygen in a wetland soils (Lovley *et al.*,, 1996; Keller *et al.*, 2009). In fact, anaerobic microorganisms could reduce humic substances or natural organic matter (NOM), which then donates electrons to reduce Fe (III)-containing minerals to release soluble $Fe^{2+}$. In anaerobic environment at low pH, abiotic reduction of $Fe^{3+}$ by NOM is well reported (Chen *et al.*, 2003). Above mentioned processes not only accelerate the chemical reduction of minute quantities of $Fe^{3+}$ species in solutions but also facilitate the reduction of solid phase ferric iron minerals.

So, quantity, nature and mode of decomposition (aerobic or anaerobic) of the organic matter are the most important factors to determine the redox status and redox reactions in the soil.

***Microorganisms:*** The microorganisms of the soil can be divided into three groups according to their oxygen needs:

aerobic, facultative anaerobic and obligate anaerobic. The aerobic bacteria use only oxygen, facultative anaerobic can use both oxygen and other elements while obligate anaerobic can use only other elements as Terminal Electron Acceptors (TEAs). So, activities of all types of microorganisms are affected directly by oxygen level and in turn they affect the redox status of the soil. The microbial activity is, indirectly, influenced by many factors such as organic matter availability and decomposability, temperature, carbon dioxide concentration, soil bulk density, soil pH, soil porosity, and soil water contents (Linn *et al.*, 1984; Glinski and Stepniewski, 1985).

There is a large diversity in the aerobic and anaerobic bacteria. Some are particularly involved in the oxidation or reduction of a specific element while others have the ability to utilize a large range of alternative TEAs (Losekann *et al.*, 2007). For example, $Fe^{3+}$ reducing bacteria are normally in active colonies and are capable of using other elements as TEAs (Lovley, 2000). Iron reducing bacteria have been found in almost every place in the Earth. *Geobacter spp*, and *Shewanella spp*. are Fe reducing bacteria and are commonly found in saline environments while *Thiobacillus ferrooxidans* and *Leptospirillum ferrooxidans* are common Fe oxidizing bacteria (Lovley, 2000).

On the basis of water requirements, microbes can be differentiated as hygrophiles, mesophiles and xerophiles. Hygrophiles include majority of bacteria, yeasts and certain fungi, majority of fungi belong to mesophile class while xerophiles consist of *Phycomycetes spp*. (Glinski and Stepniewski, 1985). Water availability can be a limitation factor for the microbial activity. A low soil respiration is observed at low soil moisture or at high soil moisture due to low water availability or low oxygen availability respectively (Orchard and Cook, 1983) and soil respiration comes to stand still at a soil water potential above 10 MPa (Glinski and Stepniewski, 1985).

Temperature is also an important factor affecting the activity and respiration of microbes (Andersson and Nilsson, 2001). So, any change in temperature can cause change in microbial activity and redox potentials (as in Nernst equation). Microbes on the basis of temperature requirements are classified as cryophiles, mesophiles and thermophiles. Optimum temperature for cryophile microbes is less than 20 °C, for mesophiles is between 20 and 30 °C while for thermophile optimal temperature is more than 40 °C (Glinski and Stepniewski, 1985; Lovley, 2000). Soil respiration and microbial activity, in general, increase with the increase in temperature but maximum soil respiration occurs between 40 and 70 °C (Andersson and Nilsson, 2001; Vepraskas and Faulkner, 2001).

***Effects of redox reactions on the soil:*** Redox conditions influence directly on the soil morphology, nutrients availability, and metals and pollutants mobility or/and accumulation in the soil (Doyle and Otte, 1997). So ultimately, redox reactions can result in change of soil fertility and increase in soil pH owing to denitrification and reduction of $Fe^{3+}$ to $Fe^{2+}$ respectively (Ponnamperuma, 1985). Oxidation conditions in the soil result in the accumulation of iron in the soil in the form of $Fe^{3+}$ while reducing conditions can result in dissolution and removal of iron from soil horizon in the form of $Fe^{2+}$. This depletion or accumulation of iron in the soil can range from a ped of soil to landscape level. Same is the case with other elements used as terminal electron acceptors (i.e. $Mn^{4+}$) in the redox reactions. As iron contents in most soils are larger than nitrate and manganese, the availability of reducible $Fe^{3+}$ induces the buffering capacity which protects soil against a rapid decrease in redox potential and ultimately mobilization of toxic ions in the soil (Förstner *et al.*, 1994; Lovley, 2000).

Reduced conditions can result in dissolutions of minerals and as well as in their precipitations; sulphates can be reduced to sulphide that can react with transition metals and precipitate in the form of insoluble metallic sulphide (i.e. FeS) (Gao *et al.*, 2004). Presence of iron minerals, such as lepidocrocite, that indicate hydromorphism can be observed in compacted soils by naked eye (Herbauts *et al.*, 1996). In one experiment Herbauts *et al.* (1996) reported the larger concentrations of easily reducible iron $Fe^{2+}$ in above 30 cm of soils after the forest soil compaction due to mechanical logging. As activity of $Fe^{2+}$ increases with the decrease in the Eh, so, $Fe^{2+}$ concentrations in the soil can be a good indicator of reductive conditions of soil (Schaller *et al.*, 1997; Ratering and Schnell, 2000).

Hydromorphic conditions, due to flooding or soil compaction, in the soils can result in the lower redox potentials (Herbauts *et al.*, 1996) and increased concentrations of Ca, Mg, K, Mn and Fe in the calcareous soils while decreased concentrations of $NO_3^-$ and P (Larson *et al.*, 1991; Phillips and Greenway, 1998). Increased solubility of cations can be due to increased solubility of dissolved organic carbon (Wolt, 1994) or may be attributed to increased competition between the cations for negative charged sites (Phillips and Greenway, 1998) as increase in $Fe^{2+}$ and $Mn^{2+}$ concentrations are many fold than other cations under reducing conditions (Larson *et al.*, 1991). These changes are more evident in organic matter rich soils as the exchange/adsorption sites for cations decrease when the solubility of organic matter increases (Phillips and Greenway, 1998).

***Conclusion:*** Hydromorphic soils are particular types of soils that are characterized by oxido-reduction processes (redox reactions) and these soils are widely distributed throughout the world. Redox reactions are directly affected by the level of oxygen, form and concentration of organic matter, and type and population of microorganisms. The redox reactions in the hydromorphic soils largely influence the availability of macro and micro nutrients, dissolution or precipitation of several minerals, and, sometimes, release of toxic heavy metals in the soil solution. There is great need of research to understand these complex and dynamic processes for efficient utilization and conservation of hydromorphic lands.

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## REFERENCES

Andersson, S. and S. Nilsson. 2001. Influence of pH and temperature on microbial activity, substrate availability of soil-solution bacteria and leaching of dissolved organic carbon in a mor humus. Soil Biol. Biochem. 33: 1181-1191.

Bijlsma, L. 1996. Coastal zones and small islands. Cambridge University Press, New York, NY(USA). 289-324.

Bohn, H., B. Mcneal and G. O. Connor. 1985. Oxidation and Reduction, Soil Chemistry, John Wiley & sons New York.

Bokhari, S. 1980. Severity of waterlogging and salinity problems in Pakistan. Ministry of Water and Power Pakistan 440: 28.

Bouwman, L. and W. Arts. 2000. Effects of soil compaction on the relationships between nematodes, grass production and soil physical properties. Appl. Soil Ecol. 14: 213-222.

Brzezinska, M., Z. St pniewska and W. St pniewski. 1998. Soil oxygen status and dehydrogenase activity. Soil Biol. Biochem. 30: 1783-1790.

Cannell, R. 1977. Soil aeration and compaction in relation to root growth and soil management. Appl. Boil. 2: 1-86.

Chen, J., B. Gu, R. Royer and W. Burgos. 2003. The roles of natural organic matter in chemical and microbial reduction of ferric iron. Sci. Total Environ. 307: 167-178.

Czy , E. 2004. Effects of traffic on soil aeration, bulk density and growth of spring barley. Soil Till. Res. 79: 153-166.

Dobos, R., E. Ciolkosz and W. Waltman. 1990. The effect of organic carbon, temperature, time, and redox conditions on soil color. Soil Sci. 150: 506-512.

Doyle, M. and M. Otte. 1997. Organism-induced accumulation of iron, zinc and arsenic in wetland soils. Environ. Pollut. 96: 1-11.

Drew, M. 1990. Sensing soil oxygen. Plant Cell Environ. 13: 681-693.

Duchaufour, P. 1997. Abrégé de pédologie. Sol, vegetation, environnement. Masson.

Fiedler, S., M. Vepraskas and J. Richardson. 2007. Soil redox potential: importance, field measurements, and observations. Adv. Agron. 94: 1-54.

Förstner, U., W. Calmano, J. Hong and M. Kersten. 1994. Effects of Redox Variations on Metal Speciation--Implications on Sediment Quality Criteria Assessment. Sp. Pub. Royal Soc. Chem. 154: 83-102.

Gao, S., K. Tanji and S. Scardaci. 2004. Impact of rice straw incorporation on soil redox status and sulfide toxicity. Agron. J. 96: 70-76.

Glinski, J. and W. Stepniewski. 1985. Soil aeration and its role for plants, Fl., CRC Press, Boca Raton.

Grable, A. and E. Siemer. 1968. Effects of bulk density, aggregate size, and soil water suction on oxygen diffusion, redox potentials, and elongation of corn roots. Soil Sci. Soc. Am. J. 32: 180-186.

Hemmat, A. and O. Taki. 2003. Comparison of compaction and puddling as pre-planting soil preparation for mechanized rice transplanting in very gravelly Calcisols in central Iran. Soil Till. Res. 70: 65-72.

Herbauts, J., J. El Bayad and W. Gruber. 1996. Influence of logging traffic on the hydromorphic degradation of acid forest soils developed on loessic loam in middle Belgium. Forest Ecol. Manag. 87: 193-207.

Hillel, D. 1980. Fundamentals of soil physics, Academic Press New York.

Jacob, D. and M. Otte. 2003. Conflicting processes in the wetland plant rhizosphere: Metal retention or mobilization? Water Air Soil Poll.: Focus 3: 91-104.

Keller, J., P. Weisenhorn and J. Megonigal. 2009. Humic acids as electron acceptors in wetland decomposition. Soil Biol. Biochem. 41: 1518-1522.

Khan, S. R. A. 2010. "Water resource planning in Tenth Five Year Plan, http://www.pakissan.com/english/watercrisis/water.resource.planning.in.tenth.five.year.plan.shtml visted on 23 August 2013."

Kowalchuk, G., P. Bodelier, G. Heilig, J. Stephen and H. Laanbroek. 1998. Community analysis of ammonia-oxidising bacteria, in relation to oxygen availability in soils and root-oxygenated sediments, using PCR, DGGE and oligonucleotide probe hybridisation. FEMS Microbiol. Ecol. 27: 339-350.

Larson, K., D. Graetz and B. Schaffer. 1991. Flood-induced chemical transformations in calcareous agricultural soils of south Florida. Soil Sci. 152: 33-40.

Linn, D., J. Doran and A. USDA. 1984. Aerobic and anaerobic microbial populations in no-till and plowed soils. Soil Sci. Soc. Am. J. 48: 794-799.

Losekann, T., K. Knittel, T. Nadalig, B. Fuchs, H. Niemann, A. Boetius and R. Amann. 2007. Diversity and abundance of aerobic and anaerobic methane oxidizers at the Haakon Mosby Mud Volcano, Barents Sea. Appl. Environ. Microbiol. 73: 3348-3362.

Lovley, D. 1987. Organic matter mineralization with the reduction of ferric iron: a review. Geomicrobiol. J. 5: 375-399.

Lovley, D. 2000. Environmental microbe-metal interactions. Washington, DC, ASM Press.

Lovley, D., J. Coates, E. Blunt-Harris, E. Phillips and J. Woodward. 1996. Humic substances as electron acceptors for microbial respiration. Nature 382: 445-448.

McBride, M. 1994. Environmental chemistry of soils, Oxford University Press New York.

Megonigal, J., M. Hines and P. Visscher. 2004. Anaerobic metabolism: linkages to trace gases and aerobic metabolism. Biogeochemistry: 317–424.

Murtaza, B., G. Murtaza, M. Z. Rehman and A. Ghafoor. 2011. Use of brackish water for the reclamation of dense saline sodic soils by auger hole technology and economic growth of wheat. Pak. J. Agri. Sci. 48: 269-276.

Nawaz, M. F., G. Bourrie and F. Trolard. 2013. Soil compaction impact and modelling. A review. Agron. Sustain. Dev. 33: 291-309.

Nawaz, S., R. Qureshi, M. Aslam and S. Parveen. 1992. Comparative performance of different wheat varieties under salinity and waterlogging. Pak. J. Agri. Sci. 29: 376-380.

Orchard, V. and F. Cook. 1983. Relationship between soil respiration and soil moisture. Soil Biol. Biochem. 15: 447-453.

Ouyang, Y. and L. Boersma. 1992. Dynamic oxygen and carbon dioxide exchange between soil and atmosphere: I. Model development. Soil Sci. Soc. Am. J. 56: 1695-1702.

Panayiotopoulos, K.,C.Papadopoulou and A. Hatjiioannidou. 1994. Compaction and penetration resistance of an Alfisol and Entisol and their influence on root growth of maize seedlings. Soil Till. Res. 31: 323-337.

Phillips, I. and M. Greenway.1998.Changes in water-soluble and exchangeable ions, cation exchange capacity and phosphorus sorption capacity under alternate waterlogged and drying conditions. Commun. Soil Sci. Plant Anal 29: 51–65.

Ponnamperuma, F. 1972. The chemistry of submerged soils. Adv. Agron. 24: 29-96.

Ponnamperuma, F. 1985. Chemical kinetics of wetland rice soils relative to soil fertility. Wetland Soils, characterization, Classification and Utilization, Philippine, Agribookstore.

Qureshi, A., P. McCornick, M. Qadir and Z. Aslam. 2008. Managing salinity and waterlogging in the Indus Basin of Pakistan. Agri. Water Manag. 95: 1-10.

Ratering, S. and S. Schnell. 2000. Localization of iron-reducing activity in paddy soil by profile studies. Biogeochemistry 48: 341-365.

Schaller, T., H. Christoph Moor and B. Wehrli. 1997. Sedimentary profiles of Fe, Mn, V, Cr, As and Mo as indicators of benthic redox conditions in Baldeggersee. Aquat. Sci.-Res. Acr. Bound. 59: 345-361.

Schüring, J., H. Schulz, W. Fischer and J. Böttcher. 2000. Redox: fundamentals, processes, and applications, Springer Verlag.

Silver, W., A. Lugo and M. Keller. 1999. Soil oxygen availability and biogeochemistry along rainfall and topographic gradients in upland wet tropical forest soils. Biogeochemistry 44: 301-328.

Sparks, D. 2003. Redox Chemistry of Soils, Environmental soil chemistry, Academic Pr.

Vepraskas, M. and S. Faulkner. 2001. Redox chemistry of hydric soils. Wetland soils: Genesis, hydrology, landscapes, and classification. J. Richardson and M. Vepraskas. Washington, D.C., Lewis Publisher: 85–106.

Wolt, J. 1994. Soil solution chemistry: applications to environmental science and agriculture, John Wiley & Sons New York, NY, USA.

WSR. 2001. World Soil Resources Reports. Lecture notes on the major soils of the world. FAO United Nations, Rome: 115-121.

WSR. 2006. World Soil Resources Reports 103. World reference base for soil resources. Rome: 80-81.