Technical Session I :

Potassium in Soils

Dynamics of K in Soils and Their Role in Management of K Nutrition

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Introduction

The role of K in soils is prodigious. Of the many plant nutrient-soil mineral relationships, those involving K are of major if not prime significance (Sparks and Huang, 1985). Since the middle of the 17th century, when J.R. Glauker in The Netherlands first proposed that saltpeter (KNO_3) was the principle of vegetation, K has been recognized as being beneficial to plant growth (Russell, 1961). Glauker obtained large increases in plant growth from addition of saltpeter to the soil that was derived from the leaching of coral soils. The essentiality of K to plant growth has been known since the work of von Liebig published in 1840 (Sparks, 2000).

Of the major nutrient elements, K is usually the most abundant in soils (Reitemeier, 1951). Igneous rocks of the Earth's crust have higher K contents than sedimentary rocks. Of the igneous rocks, granites and syenites contain 46 to 54, basalts 7, and peridotites 2.0 g K kg^{-1}. Among the sedimentary rocks, clayey shales contain 30, whereas limestones have an average of only 6 g kg^{-1} (Malavolta, 1985). Mineral soils generally range between 0.04 and 3% K. Total K contents in soils range between 3000 and 100,000 kg ha^{-1} in the upper 0.2 m of the soil profile. Of this total K content, 98% is bound in the mineral form, whereas 2% is in soil solution and exchangeable phases (Schroeder, 1979; Bertsch and Thomas, 1985).

Potassium, among mineral cations required by plants, is the largest in nonhydrated size (r = 0.133 nm) and the number of oxygen atoms surrounding it in mineral structures is high (8 or 12), which suggests that the strength of each K-O bond is relatively weak (Sparks and Huang, 1985). Potassium has a polarizability equal to 0.088 nm^3, which is higher than for Ca^{2+}, Li^+, Mg^{2+}, and Na^+ but lower than for Ba^{2+}, Cs^+, NH^+_4, and Rb^+ ions (Rich, 1968, 1972; Sparks and Huang, 1985). Ions with higher polarizability are preferred in ion exchange reactions. Potassium has a hydration energy of 142.5 kJ g^{-1} ion^{-1}, which indicates little ability to cause soil swelling (Helfferich, 1962).

Forms of Soil K

Soil K exists in four forms in soils: solution, exchangeable, fixed or nonexchangeable, and structural or mineral (**Figure 1**). Quantities of exchangeable, nonexchangeable, and total K in the surface layer (0-20 cm) of a variety of soils are shown in **Table 1**. Exchangeable K and nonexchangeable K levels comprise a small portion of the total K. The bulk of total soil K is in the mineral fraction (Sparks and Huang, 1985). There are equilibrium and kinetic reactions between the four forms of soil K that affect the level of soil solution K at any particular time, and thus, the amount of readily available K for plants. The forms of soil K in the order of their availability to plants and microbes are solution > exchangeable > fixed (nonexchangeable) > mineral (Sparks and Huang, 1985; Sparks, 1987; Sparks, 2000).

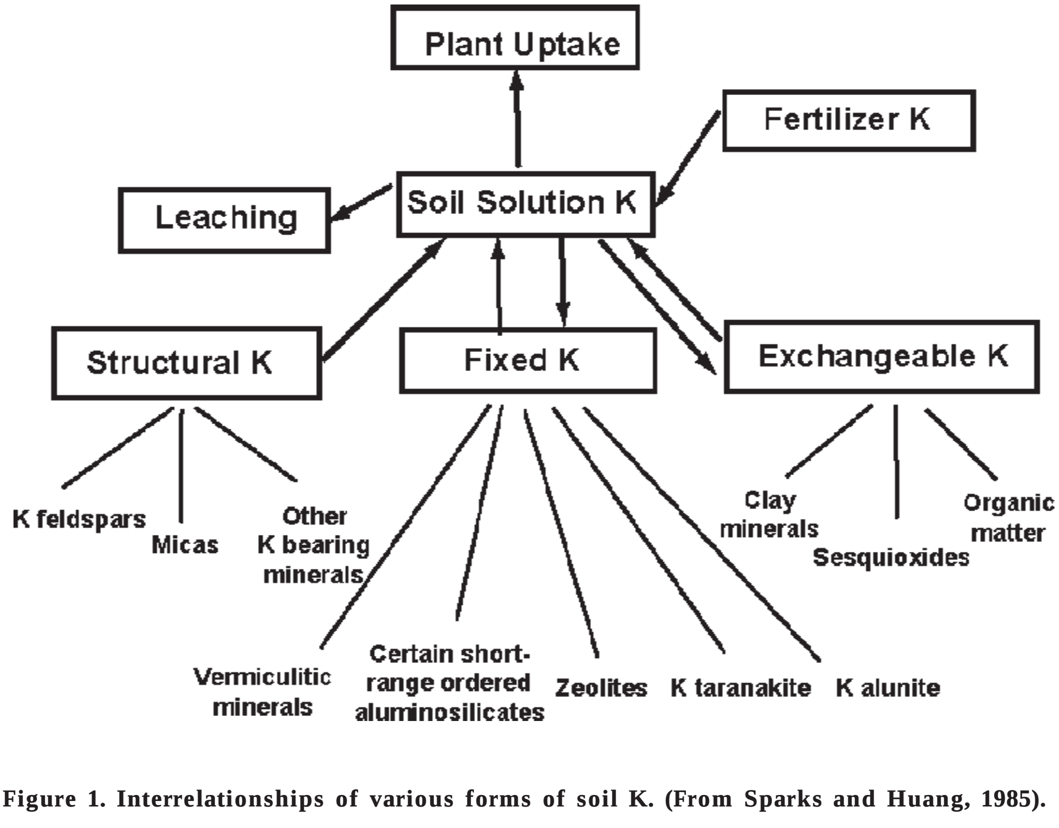

Figure 1. Interrelationships of various forms of soil K. (From Sparks and Huang, 1985).

Table 1. Potassium status of selected soils.

Origin of Soil	***Exchangeable K***	***Nonexchangeable K***	***Total K***	***Source***
	(cmol kg^{-1})			
Alfisols				
Nebraska, USA	0.40	--	--	*Soil Taxonomy* (1975)
West Africa	0.46	--	3.07	Juo (1981)
Inceptisols				
California, USA	0.40	--	--	*Soil Taxonomy* (1975)
Maryland, USA	0.20	--	--	*Soil Taxonomy* (1975)
Mollisols				
Iowa, USA	0.27	--	--	*Soil Taxonomy* (1975)
Nebraska, USA	0.40	--	--	
Ultisols				
Delaware, USA	0.33	0.49	22.5	Parker *et al.* (1989)
Florida, USA	0.14	0.25	2.71	Yuan *et al.* (1976)
Virginia, USA	0.11	0.17	6.5	Sparks *et al.* (1980)
West Africa	0.24	--	8.06	Juo (1981)

a Data are for surface soils (0-20 cm depth).

Soil solution K is the form of K that is directly taken up by plants and microbes and also is the form most subject to leaching in soils. Levels of soil solution K are generally low, unless recent amendments of K have been made to the soil. The quantity of K in the soil solution varies from 2 to 5 mg K L^{-1} for normal agricultural soils of humid regions and is an order of magnitude higher in arid region soils (Haby *et al.*, 1990). Levels of solution K are affected by the equilibrium and kinetic reactions that occur between the forms of soil K, the soil moisture content, and the concentrations of bivalent cations in solution and on the exchanger phase (Sparks and Huang, 1985; Sparks, 2000).

Exchangeable K

Exchangeable K is the portion of the soil K that is electrostatically bound as an outer-sphere complex to the surfaces of clay minerals and humic substances. It is readily exchanged with other cations and also is readily available to plants.

Nonexchangeable K

Nonexchangeable or fixed K differs from mineral K in that it is not bonded within the crystal structures of soil mineral particles. It is held between adjacent tetrahedral layers of dioctahedral and trioctahedral micas, vermiculites, and intergrade clay minerals such as chloritized vermiculite (Rich, 1972; Sparks and Huang, 1985; Sparks, 1987). Potassium becomes fixed because the binding forces between K and the clay surfaces are greater than the hydration forces between individual K^+ ions. This results in a partial collapse of the crystal structures and the K^+ ions are physically trapped to varying degrees, making K release a slow, diffusion controlled process (Sparks, 1987). Nonexchangeable K also can be found in wedge zones of weathered micas and vermiculites (Rich, 1964). Only ions with a size similar to K^+, such as NH_4^+ and H_3O^+, can exchange K from wedge zones. Large hydrated cations, such as Ca^{2+} and Mg^{2+}, cannot fit into the wedge zones. Release of nonexchangeable K to the exchangeable form occurs when levels of exchangeable and soil solution K are decreased by crop removal and/or leaching and perhaps by large increases in microbial activity (Sparks, 1980; Sparks, 2000).

Nonexchangeable K is moderately to sparingly available to plants (Mengel, 1985; Sparks and Huang, 1985; Sparks, 1987). Mortland *et al.* (1956) showed that biotite could be altered to vermiculite by plant removal of K. Schroeder and Dummler (1966) showed that the nonexchangeable K associated with some German soil illites was an important source of K to crops. The ability of plants to take up nonexchangeable K appears to be related to the plant species. Steffens and Mengel (1979) found that ryegrass (*Lolium perenne*) could take up nonexchangeable K longer without yield reductions, while red clover (*Trifolium pratense*) could not. This was attributed to the ryegrass having longer root length which would allow it to grow at a relatively low K concentration. A similar concentration would result in a K deficiency in red clover. It may be that the difference in root mass, root length, and root morphology between monocots and dicots explains why monocots feed better from nonexchangeable K than dicots (Mengel, 1985).

Mineral K

As noted earlier, most of the total K in soils is in the mineral form, mainly as K-bearing primary minerals such as muscovite, biotite, and feldspars. For example, in some Delaware soils, Sadusky *et al.* (1987) found that mineral K comprised about 98% of the total K (**Table 2**). Most of the mineral K was present as K feldspars in the sand fractions.

Table 2. Potassium status of Delaware soils and sand fractions[a].

Horizon	*Depth*	*Soils*			*Sand fractions*		
		$CaCl_2$ Extractable	*HNO_3 Extractable*	*Mineral K[b]*	*Total K*	*Total K[c]*	*K feldspars[d]*
	cm	*cmol kg^{-1}*					*Frequency %*
Kenansville loamy sand							
Ap	0-23	0.25	0.42	35.02	35.69	30.88	9.5
Bt2	85-118	0.25	0.49	45.30	46.04	33.86	12.0
Rumford loamy sand							
Ap	0-25	0.33	0.49	21.67	22.51	18.62	6.7
BC	89-109	0.21	0.54	23.39	23.96	16.76	8.2
Sassafras fine loamy sand							
Ap	0-20	0.35	0.56	43.54	44.45	28.95	16.0
Cl	84-99	0.13	0.36	45.99	46.68	36.69	24.0

[a] From Sandusky *et al.* (1987).

[b] Mineral K = [(total K) - ($CaCl_2$ ext. K + HNO_3 ext. K)].

[c] These data represent the amount of total K in the sand based on a whole soil basis.

[d] Determined through petrographic analyses of the whole sand fractions and represents the percentage of total point counts in a given sample that were K feldspars. The remaining minerals in the sand fractions were quartz, plagioclase, and opaques.

Common soil K-bearing minerals, in the order of availability of their K to plants, are biotite, muscovite, orthoclase, and microcline (Huang *et al.*, 1968; Sparks, 1987). Mineral K is generally assumed to be only slowly available to plants; however, the availability is dependent on the level of K in the other forms, and the degree of weathering of the feldspars and micas constituting the mineral K fraction (Sparks and Huang, 1985; Sparks, 1987). Sadusky *et al.* (1987) and Parker *et al.* (1989a, b) have found that a substantial amount of K is released from the sand fractions of Delaware soils that are high in K feldspars. This finding, along with the large quantities of mineral K in these and other Atlantic Coastal Plain soils, could help in explaining the often observed lack of crop response to K amendments on these soils (Liebhardt *et al.*, 1976; Yuan *et al.*, 1976; Sparks *et al.*, 1980; Woodruff and Parks, 1980;

Parker *et al.*, 1989b).

Factors Affecting Potassium Availability

Solution-Exchangeable K Dynamics

The rate and direction of reactions between solution and exchangeable forms of K determine whether applied K will be leached into lower horizons, taken up by plants, converted into unavailable forms, or released into available forms (**Figure 2**).

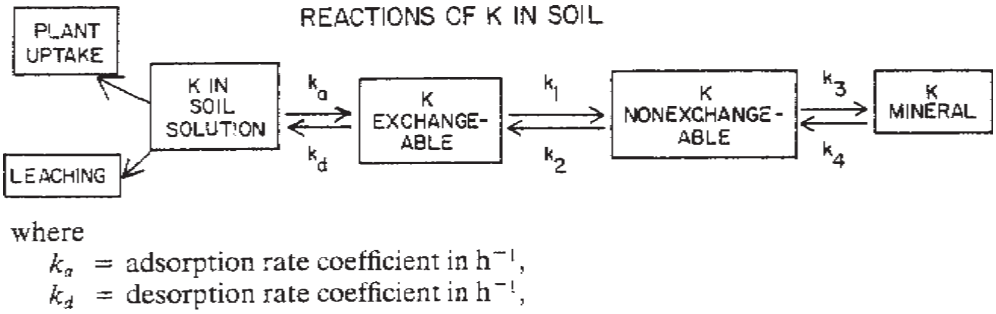

Figure 2. Kinetic reactions between soil K phases. (From Selim *et al.*, 1976).

The reaction rate between soil solution and exchangeable phases of K is strongly dependent on the type of clay minerals present (Sivasubramaniam and Talibudeen, 1972; Sparks *et al.*, 1980; Sparks and Jardine, 1981, 1984; Jardine and Sparks, 1984) and the method employed to measure kinetics of K exchange (Sparks, 1989, 1995; Amacher, 1991; Sparks *et al.*, 1996). Vermiculite, montmorillonite, kaolinite, and hydrous mica vary drastically in their ionic preferences, ion binding affinities, and types of ion exchange reactions. Such fundamental differences in these clay minerals account for the varying kinetics of K exchange.

Kinetics of K exchange on kaolinite and montmorillonite are usually quite rapid (Malcolm and Kennedy, 1969; Sparks and Jardine, 1984). An illustration of this is shown in **Figure 3**. In the case of kaolin clays, the tetrahedral layers of adjacent clay layers are held tightly by H bonds; thus, only planar external surface and edge sites are available for ionic exchange. With montmorillonite, the inner peripheral space is not held together by H bonds, but instead is

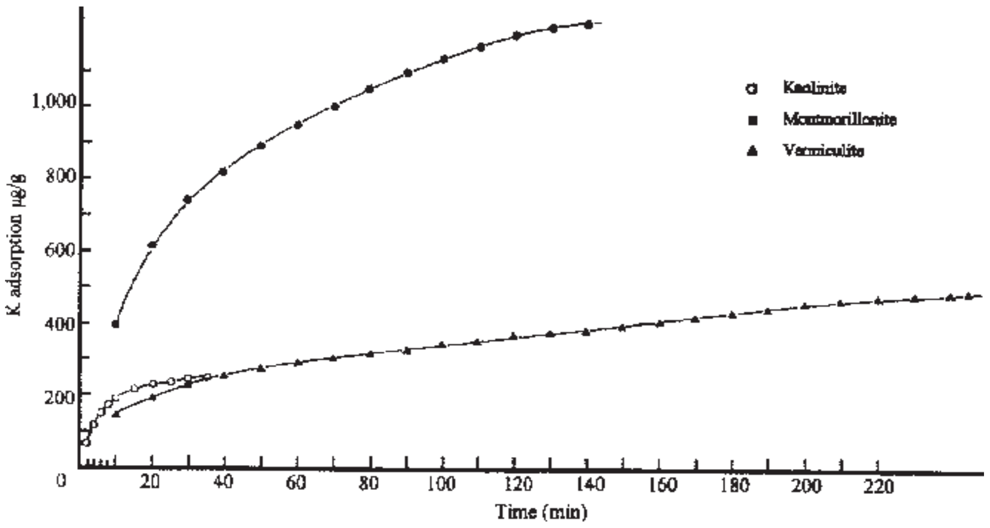

Figure 3. Potassium adsorption versus time in pure systems. (From Sparks and Jardine, 1984).

able to swell with adequate hydration, and thus allow for rapid passage of ions into the interlayer space. Malcolm and Kennedy (1969) found that the rate of Ba exchange on kaolinite and montmorillonite was rapid with 75% of the total exchange occurring in three seconds.

Kinetics of K exchange on vermiculitic and micaceous minerals tend to be extremely slow. Both are 2 : 1 phyllosilicates with peripheral spaces that impede many ion exchange reactions. Micaceous minerals typically have a more restrictive interlayer space than vermiculite since the area between layer silicates of the former is selective for certain types of cations (e.g., K^+, Cs^+). Bolt *et al.* (1963) theorized the existence of three types of binding sites for K exchange on a hydrous mica. The authors hypothesized that slow kinetics were due to internal exchange sites, rapid kinetics to external planar sites, and intermediate kinetics to edge sites.

Rates of K Exchange in Soils

A number of studies have appeared in the literature on kinetics of K exchange in soils (see references in Sparks and Huang, 1985; Sparks, 1987; Sparks, 2000). Sparks and Jardine (1981) studied K adsorption/desorption on a Matapeake soil from Delaware. Potassium desorption was slower than adsorption. This was ascribed to the difficulty in desorbing K from partially collapsed interlayer sites of the vermiculitic clay minerals. Jardine and Sparks (1984) described the kinetics of K adsorption / desorption on an Evesboro soil with two simultaneous first-order equations at 283 and 298K. These reactions were ascribed to sites of varying K reactivity. Reaction 1, the more rapid one that terminated in 10 to 12 minutes was ascribed to external surface sites of the organic and inorganic phases of soil that are readily accessible for exchange. Reaction 2, that continued for many hours, was attributed to interlayer sites of the 2 : 1 clay minerals that predominated in the < 2 mm clay fraction.

Potassium Fixation

The phenomenon of K fixation or retention significantly affects K availability. The fact that fixation processes are limited to interlayer ions such as K^+ has been explained in terms of the good fit of K^+ ions (the crystalline radius and coordination number are ideal) in an area created by holes and adjacent oxygen layers (Barshad, 1951). The important forces involved in interlayer reactions in clays are electrostatic attractions between the negatively charged layers and the positive interlayer ions, and expansive forces due to ion hydration (Kittrick, 1966).

The degree of K fixation in clays and soils depends on the type of clay mineral and its charge density, the degree of interlayering, the moisture content, the concentration of K^+ ions as well as the concentration of competing cations, and the pH of the ambient solution bathing the clay or soil (Rich, 1968; Sparks and Huang, 1985). The major clay minerals responsible for K fixation are montmorillonite, vermiculite, and weathered micas. In acid soils, the principal clay mineral responsible for K fixation is dioctahedral vermiculite. Weathered micas fix K under moist as well as dry conditions, whereas some mont-morillonites fix K only under dry conditions (Rich, 1968). The degree of K fixation is strongly influenced by the charge density on the layer silicate. Those with high charge density fix more K than those with low charge density (Walker, 1957). Weir (1965) noted that K fixation by montmorillonites is limited unless the charge density of the clays is high. Low charge montmorillonite (Wyoming) stays at 1.5 nm when K saturated unless it is heated (Laffer *et al.*, 1966). Schwertmann (1962a, b) noted that soil montmorillonites have a greater capacity to fix K than do many specimen montmorillonites. Soil montmorillonites have higher charge density and a greater probability of having wedge positions near mica-like zones where the selectivity for K is high (Rich, 1968).

The importance of interlayer hydroxy Al and hydroxy Fe^{3+} material on K fixation was first noted in the classic work of Rich and Obenshain (1955). They theorized that hydroxy Al and hydroxy Fe^{3+} interlayer groups acted as props to decrease K fixation. This theory was later corroborated in the work of Rich and Black (1964) who found that the introduction of hydroxy Al groups into Libby vermiculite increased the Gapon selectivity coefficient (kG) from 5.7 to 11.1 x 10^{-2} L $mol^{-1/2}$.

Wetting and drying and freezing and thawing can significantly affect K fixation (Hanway and Scott, 1957; McLean and Simon, 1958; Cook and Hutcheson, 1960). The degree of K fixation or release on wetting or drying is dependent on the type of colloid present and the level of K^+ ions in the soil solution. Potassium fixation by 2 : 1 clay minerals may be strongly influenced by the kind of adsorbed cations or the anions within the system. In studies with the silicate ion, Mortland and Gieseking (1951) found that montmorillonite clays dried with K_2SiO_3 were altered in their swelling properties and fixed K in large amounts. Hydrous mica clays also fixed large amounts of K that could not be removed with boiling HNO_3.

Volk (1934) observed a marked increase in K fixation in soils where pH was raised to about 9 or 10 with Na_2CO_3. Martin *et al.* (1946) showed at pH values up to 2.5 there was no fixation; between pH 2.5 and 5.5, the amount of K fixation increased very rapidly. Above pH 5.5, fixation increased more slowly. These differences in K fixation with pH were discussed by Thomas and Hipp (1968). At pH values > 5.5, Al^{3+} cations precipitate as hydroxy polycations, which increase in the number of OH groups as pH increases until they have a form like gibbsite (Thomas, 1960). At this pH (~8), Al^{3+} does not neutralize the charge on the clay and cannot prevent K fixation. Below pH 5.5, Al^{3+} and $Al(OH)_x$ species dominate. Below pH 3.5, H_3O^+ predominates (Coleman and Harward, 1953; Thomas and Hipp, 1968).

The increase in K fixation between pH 5.5 and 7.0 can be ascribed to the decreased numbers of $Al(OH)_x$ species which decrease K fixation (Rich and Obenshain, 1955; Rich, 1960, 1964; Rich and Black, 1964). At low pH, the lack of K fixation is probably due to large numbers of H_3O^+ and their ability to replace K as well (Rich, 1964; Rich and Black, 1964).

Potassium Release from Soil Minerals

The release of K from micas proceeds by two processes: (1) the transformation of K-bearing micas to expansible 2 : 1 layer silicates by exchanging the K with hydrated cations, and (2) the dissolution of the micas followed by the formation of weathering products. The relative importance of these two mechanisms depends on the stability of micas and the nature of soil environments (Sparks and Huang, 1985; Sparks, 2000).

Release of K from feldspars appears to involve a rapid exchange with H, which creates a thin layer of hydrolyzed aluminosilicates. This residual layer ranges in thickness from several to a few tens of nm and seems to cause the initial nonstoichiometric release of alkali and alkaline earths relative to Si and Al. Following this step, there is continued dissolution, which removes hyperfine particles. After these are removed, further dissolution breaks down the outer surface of the residual layer at the same rate that alkalis are replaced by H at the interface between fresh mineral surfaces and the residual layer. This releases all constituents to the solution. Release is now stoichiometric. Thus, the weathering of feldspars appears to be a surface-controlled reaction (Sparks, 1989).

A number of physiochemical and mineralogical factors govern the release of K from micas by both cation exchange reactions and dissolution processes. These include tetrahedral rotation and cell dimensions, degree of tetrahedral tilting, hydroxyl orientation, chemical composition, particle size, structural imperfections, degree of K depletion, layer charge alterations and associated reactions, hydronium ions, biological activity, inorganic cations, wetting and drying, and other factors (Sparks and Huang, 1985). This review will focus on the latter four factors.

Biological activity promotes K release from micas (Mortland *et al.*, 1956; Boyle *et al.*, 1967; Weed *et al.*, 1969; Sawhney and Voight, 1969). The organisms deplete the K in the soil solution, and their action may be analogous to that of tetraphenylboron (TPB) in artificial weathering of micas. Furthermore, the overall action of organisms is more complex when organic acids are produced (Boyle *et al.*, 1967; Spyridakis *et al.*, 1967; Sawhney and Voight, 1969).

The importance of organic acids in weathering of rock-forming minerals has been recognized for a long time (Sprengel, 1826; Bolton, 1882; Huang and Keller, 1970). All soils contain small but measurable amounts of biochemical compounds such as organic acids. Furthermore, since the time required for soil formation can extend over a period of centuries, the cumulative effect in a soil of even very small quantities of chelating agents will be considerable. The influence of oxalic and citric acids on the dynamics of K release from micas and feldspars was studied by Song and Huang (1988). They found that the sequence of K release from K-bearing minerals by oxalic and citric acids is biotite > microcline > orthoclase > muscovite.

The activity of K^+ ions in soil solution around mica particles greatly influences the release of K from micas by cation exchange. When the K level is less than the critical value, K is replaced from the interlayer by other cations from the solution. On the contrary, when the K level is greater than the critical value, the mica expansible 2 : 1 mineral takes K from the solution. The critical K level is highly mineral dependent, being much higher for the trioctahedral minerals (Scott and Smith, 1967; Newman, 1969; von Reichenbach, 1973; Henderson *et al.*, 1976). The critical levels for muscovite are so low that even the K impurities in laboratory chemicals or dissolved from glassware are often sufficient to prevent any K release (Scott and Smith, 1967).

The nature and concentration of the replacing cations also influence the critical K level of the cations tested in Cl^- solutions. Rausell-Colom *et al.* (1965) found the critical K levels decreased in the order $Ba^{2+} > Mg^{2+} > Ca^{2+} = Sr^{2+}$ for the same concentration of these ions and with a constant mica particle size. The activity of all of these replacing ions in the solution phase must be much greater than that of the K for significant K release to occur. The activity of various cations in the soil solution is governed by other minerals in the soil systems, by pedogenic processes, and by anthropogenic activities.

The release of K upon drying a soil is related to the clay fraction (Scott and Hanway, 1960). When a soil is dried, the degree of rotation of weathered soil minerals, such as micas, may be changed. Thus, the K-O bond may be modified. Dehydration of interlayer cations may permit a redistribution of interlayer cations, because Ca could now compete with K for wedge sites. This seems to account for the release of K from soils upon drying. Rich (1972) found that Virginia soils, which contain hydroxy Al interlayers and appreciable amounts of K, did not release K upon drying. The presence of the hydroxy Al interlayers may block or retard the interlayer diffusion of K ions and may change the b dimension of micas, the degree of tetrahedral rotation, and the length and strength of the K-O bond.

Other factors that can affect K release from soils are leaching, redox potential (Eh), and temperature. Leaching promotes the K release from K-bearing minerals by carrying away the reaction products. Therefore, leaching accelerates the transformation of minerals, e.g., micas, to expansible 2 : 1 layer silicates and other weathering products if the chemistry of leaching water favours the reaction. Redox potential of soils could influence K release from micas since it has been pointed out that the tenacity with which K is held by biotite is greater after oxidation of its structural Fe. It appears that, other factors being equal, the extent of the K release from biotite should be less in soil environments that oxidize Fe than in soil environments that reduce it. Major elements in K-bearing feldspars do not exist in more than one valence state, thus the prevailing Eh of a soil may not be of direct concern to chemical weathering of the feldspars. However, the weatherability of feldspars can be affected by complexing organic acids that are vulnerable to oxidation. Therefore, the stability of feldspars may be indirectly related to the prevailing Eh of a soil.

Increasing temperature has been shown to increase the rate of K release from biotite (Rausell-Colom *et al.*, 1965) and K feldspars (Rasmussen, 1972). Under conditions of leaching of biotite with 0.1 mol NaCl L^{-1}, the rate of K release appears directly proportional to temperature in the range of 293 to 323 K (Mortland, 1958). Under similar leaching conditions, Mortland and Ellis (1959) observed that the log of the rate constant for K release from fixed K in vermiculite was directly proportional to the inverse of the absolute temperature. Preheating of micas to high temperatures (1273 K) prior to TPB extraction (Scott *et al.*, 1973) was found to enhance the rate of K extraction from muscovite, to decrease the rate for biotite, and to have little effect on phlogopite, except at very high temperatures. The decrease in K release from biotite by preheating is presumably because of oxidation of Fe at high temperatures. The more rapid rate with muscovite following heating remains unexplained.

The importance of K release from soil feldspars cannot be ignored. In the past, researchers have thought that only small amounts of K from feldspars were released over a growing season to plants. However, a number of reports have appeared that show a remarkable lack of response by crops such as corn (Zea mays L.) to K applications on sandy soils (Liebhardt *et al.*, 1976; Sparks *et al.*, 1980; Woodruff and Parks, 1980; Parker *et al.*, 1989b) of the Atlantic Coastal Plain Region of the USA. The lack of crop response is anomalous, since these sandy soils have low levels of solution, exchangeable, and nonexchangeable K (Sparks, 1987; Parker *et al.*, 1989a). However, they often contain large quantities of mineral K, particularly as feldspars.

Much of the total K in soils from Delaware is contained in the sand fractions. For example, Sadusky *et al.* (1987) found that 87 and 74% of the total soil K in the Ap and B2t horizons, respectively of a Kenansville loamy sand was contained in the sand fractions. Much of this total K in the sand fractions could be directly attributable to the large quantities of K-feldspars. To assess the release rates of K from sandy soils and in particular, to determine nonexchangeable K release kinetics from feldspars, Sadusky *et al.* (1987) conducted a study on several soils from Delaware. A H-saturated resin and oxalic acid were used to affect K release. Over a 30 day period, large quantities of K were released from the soils (**Table 3**). Initially, K release increased rapidly and then began to level off as an equilibrium was approached. More K was released from the subsoil horizons than from the Ap horizons of each soil which was directly attributable to the higher clay contents in the subsoil horizons that were high in vermiculitic minerals.

Table 3. Potassium release from soils using a H-resin and oxalic acid[a,b].

Horizon	***Depth cm***	***H-Resin***	***Oxalic Acid***
		cmol kg^{-1}	
Kenansville loamy sand			
Ap	0-23	0.199	1.97×10^{-3}
Bt2	89-118	0.251	2.97×10^{-3}
Rumford loamy sand			
Ap	0-25	0.172	1.41×10^{-3}
BC	89-109	0.231	1.97×10^{-3}
Sassafras fine loamy sand			
Ap	0-20	0.235	5.64×10^{-4}
C1	84-99	0.246	1.69×10^{-3}

[a] *From Sadusky et al. (1987).*
[b] *These values represent amounts of K released at 30 d.*

Significant quantities of K were released from the coarse, medium, and fine sand fractions of the Kenansville soil to H-resin (**Table 4**) with the fine fraction releasing the most. This finding was important because the fine fraction comprised the bulk of the sand in the soils. If one averages the quality of K release from the three sand fractions, multiplies by the percentage of sand in each soil horizon, and then divides this quantity by the total K released from the whole soil, some interesting data are obtained. In the Kenansville Ap and B2t horizons, 68 and 63%, respectively of the total K released came from the sand fractions. Thus, the importance of the sand fraction in these soils in supplying K is immense. Recent research by Rahmatullah and Mengel (2000) also clearly indicates the role that sand and silt fractions of soils (**Figure 4**) play in K release.

Table 4. Potassium released after 30 d from Kenansville soil and sand fractions using H-resin[a]**.**

Whole soil		*Coarse sand*		*Medium sand*		*Fine sand*	
Horizon	*K mg kg^{-1}*	*% of whole soil*	*K released mg kg^{-1}*	*% of whole soil*	*K released mg kg^{-1}*	*% of whole soil*	*K released mg kg^{-1}*
Ap	77.5	43.0	53.6	10.0	65.3	20.5	71.6
Bt2	98.0	24.1	76.2	11.1	69.6	26.7	99.7

[a] *Adapted from Sadusky et al. (1987).*

To explain the mechanism for K release from the sand fractions, Sadusky *et al.* (1987) conducted scanning electron microscopy studies on the feldspar particles from the fine sand fractions of the soils before and after weathering with the H-saturated resin. The feldspar surfaces exhibited oval shaped and larger prismatic edge pits. Such edge pits demonstrated the highly weathered nature of the feldspars and suggested that surface-controlled kinetics was the mechanism for K release. In large part because of K release from weathered feldspars in high K containing soils, a lack of crop response to added K (**Table 5**) has often been noted for Atlantic Coastal Plain soils in the USA.

Leaching of Potassium in Soils

Soil solution K is either leached or sorbed by plants or soils. A number of factors influence the movement of K in soils, including the CEC, soil pH and liming, method and rate of K application, and K absorption by plants (Terry and McCants, 1968; Sparks, 1980; Sparks, 2000). The ability of a soil to retain applied K is very dependent on the CEC of the soil. Thus, the amount of clay and SOM in the soil strongly influences the degree of K leaching. Soils with higher CEC have a greater ability to retain added K, whereas leaching of K is often a problem on sandy soils (Sparks and Huang, 1985).

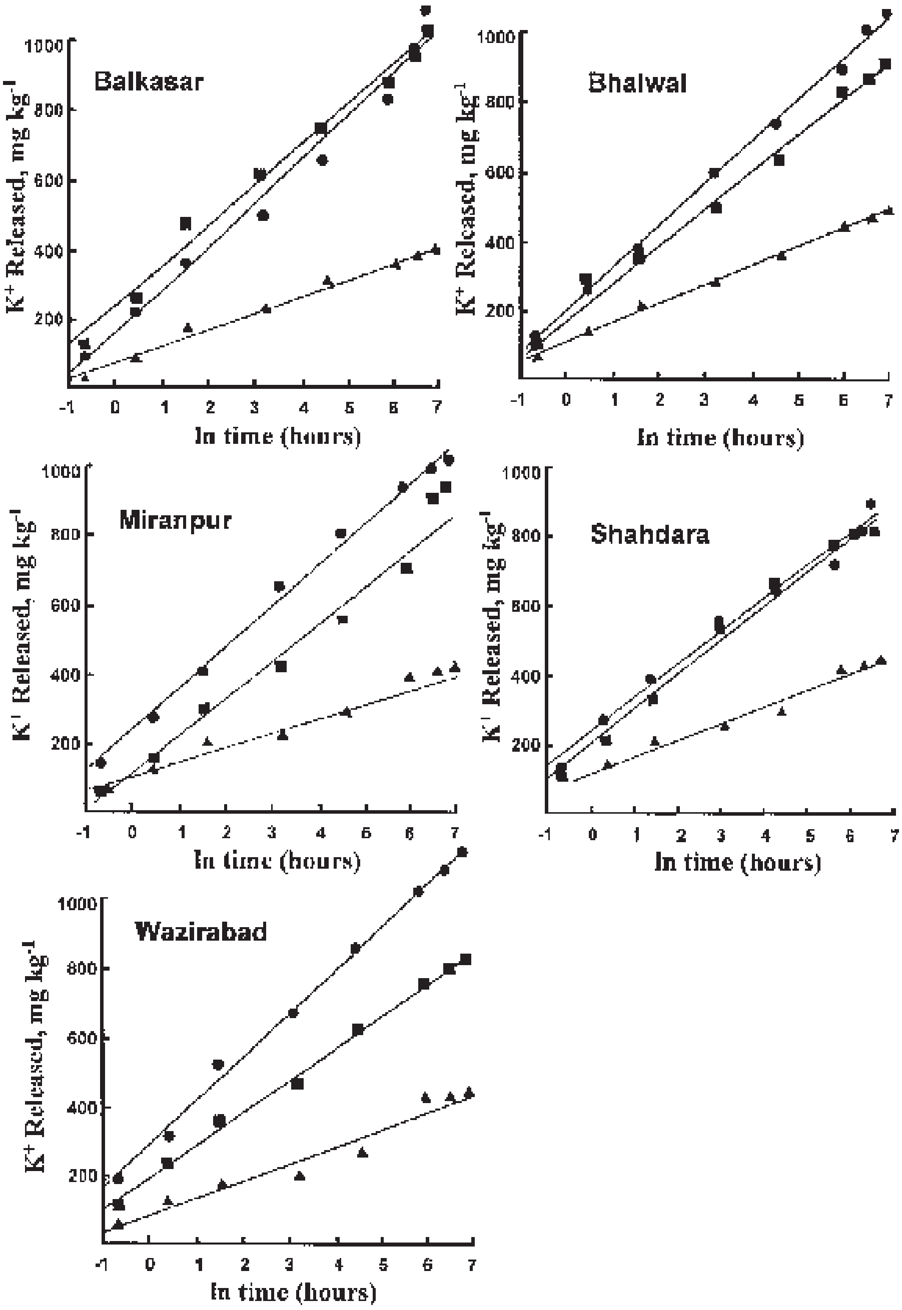

Figure 4. Release of nonexchangeable K^+ form sand (▲), silt (■) and clay (●) fractions. From Rahmatullah and Mengel (2000)

Table 5. Effect of K applications on corn grain yields at four sites over three growing seasons. Yield was not significantly ($p = 0.05$) affected by treatment for any year-site combination[a].

Annual K Application[b] kg K ha^{-1}	*Rumford*	*Kenansville*	*Matapeake*	*Sassafras*
		Soil		
	yield, Mg ha^{-1}			
		1982		
0	13.1	12.4	12.1	13.1
94	13.4	12.3	11.3	12.9
94S	14.0	12.3	11.7	13.0
282	13.5	11.4	11.1	13.1
282S	13.8	11.4	11.2	12.2
SEM[c]	0.3	0.5	0.4	0.4
		1983		
0	8.9	–	12.7	9.2
94	9.4	–	12.5	9.1
94S	8.5	–	12.1	7.6
282	11.5	–	11.8	6.9
282S	11.5	–	12.2	6.9
SEM	0.9	–	0.4	1.2
		1984		
0	9.8	–	8.7	–
94	10.1	–	9.2	–
94S	10.3	–	9.8	–
282	10.0	–	9.4	–
282S	10.5	–	9.3	–
SEM	0.2	–	0.4	–

[a] *From Parker et al. (1989b).*
[b] *S indicates K was applied in three equal portions to give the total rate indicated.*
[c] *Standard error of the mean.*

Retention of K can often be enhanced in sandy, Atlantic Coastal Plain soils after application of lime, since in such variable charge soils, the CEC is increased as soil pH is increased. Nolan and Pritchett (1960) found that liming a Lakeland fine sand soil (thermic, coated Typic Quartzipsamment) to pH 6 to 6.5 caused maximum retentivity of applied K. Potassium was replaced by Ca on the exchange complex at higher levels of limestone application. Less leaching of K occurred at pH 6.0 to 6.5 due to enhanced substitution of K for Ca than for Al, which was more abundant at low pH. Lutrick (1963) found that K leaching occurred on unlimed but not on limed areas when 112 to 224 kg K ha^{-1} was applied on a Eustis loamy fine sand (sandy, siliceous, thermic Psammentic Paleudult).

Movement of applied K has been related to the method of application. Nolan and Pritchett (1960) compared banded and broadcast placement of KCl applied at several rates to an Arredondo fine sand (loamy, siliceous, hyperthermic Grossarenic Paleudult) in lysimeters under winter and summer crops. For the low rate of application, cumulative K removal for both placements was only about 5.0 kg ha^{-1}.

A number of investigations have been conducted to determine the relationship of crop uptake and rate of K application to leaching of K. Jackson and Thomas (1960) applied up to 524 kg K ha^{-1} prior to planting sweet potatoes (*Ipomoea batatas* L.) on a Norfolk sandy loam (fine loamy, siliceous, thermic Typic Paleudult). At harvest time, soil and plant K exceeded applied K at the 131 and 262 kg K ha^{-1} rates. However, at the 524 kg K ha^{-1} rate, 38 kg K was unaccounted for by soil and plant K.

This deficiency of K was attributed to leaching below sampling depths. During a two-year study with corn (Zea mays L.) on two Dothan (Typic Paleudult) soils of Virginia, Sparks *et al.* (1980) found that 83 and 249 kg of applied K ha^{-1} increased the exchangeable K in the E and B2lt horizons of the two soils. These increases were ascribed to leaching of applied K. The magnitude of leaching varied directly with rate of K application. No accumulation of K was found in the top 0.76 m of a Leon sand (sandy, siliceous, thermic Aeric Haplaquod) after 40 yr of heavy K fertilization (Blue *et al.*, 1955).

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