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Sorption of radionuclides on geologic media — A literature survey.

I: Fission Products

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SORPTION OF RADIONUCLIDES ON GEOLOGIC MEDIA - A LITERATURE SURVEY. I: Fission Products

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ABSTRACT

A literature survey on the sorption of fission products on rocks and minerals has been performed. Some sorption data for clays and soils have also been included. The fission products investigated were cobalt, nickel, strontium, cesium, technetium and iodine. Parameters of importance to sorption have been identified and a tabulation of distribution coefficients for groundwater conditions (pH 7-9, low to medium ionic strength) is included in the report.

Some conclusions on the sorptive behaviour of these fission products may be drawn:

For cobalt and nickel the sorption is related to hydrolysis. High sorption is observed at a pH where hydrolysis becomes important and the sorption is high as long as cationic hydrolysis products are formed. When pH is increased neutral or negatively charged hydrolysis products may be formed and the sorption decreases.

For these reactions nuclide concentration CEC (cation exchange capacity) and temperature are of minor importance.

For strontium hydrolysis is of no importance at the normal pH of groundwater, but in groundwater above pH 9 carbonate complexation may occur. For most minerals, the sorption is low, usually with a pronounced pH dependence. Other important parameters are ionic strength and CEC. A nonselective sorption due to electrostatic interactions between negatively charged mineral surfaces and Sr^{2+} seems to occur.

For cesium no hydrolysis may be expected and pH has less importance than for Sr. For most minerals, however, the sorption of Cs is higher than for Sr. Important parameters are nuclide concentration and ionic strength. A selectivity for Cs-sorption is found for some minerals, mostly sheet-silicates.

For technetium sorption is due to a reduction of TcO_4 - to $TcO_2(s)$ and as anions are poorly sorbed, the sorption is dependent on the redox potential.

Iodine is also anionic and poorly sorbed. Minerals containing ions capable of forming iodides with low solubility (Ag, Mg, Pb etc) are, however, sorbing I^-.

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TABLES

1. INTRODUCTION

Measurements of the sorption of radionuclides on geologic materials started in the early fifties, mainly in the US. These early investigations had two purposes: to investigate the possibility of using natural minerals for purification of radioactive waste solutions and to study the migration potential of buried radioactive wastes in shallow land burial grounds.

Since then studies of the sorptive properties of rocks and minerals have been started all over the world, mainly in order to be able to predict the behaviour of nuclear wastes in geologic repositories.

This literature survey has been directed mainly towards the sorption of fission products on rocks and minerals, but also data for soils and clays have been included. No data measured at the department of Nuclear Chemistry at Chalmers University of Technology are included. These data are reported in separate KBS Technical Reports.

The sorption data are presented in the following text, where an evalution of parameters of importance for the sorption is included. The distribution coefficients (K_d-values) are also given in tables. When studying these tables it is important to observe that for many of the published K_d-values important data on the conditions of the measurements are not given in the publications, such as pH, nuclide concentration or ionic strength. A comparison of data from different laboratories must therefore be performed with great care.

Some of the cited data have been recalculated, usually when the distribution coefficient is not given explicitely but as % sorption or equivalent and enough data on the experimental conditions are given to allow a calculation of K_d.

2. SORPTION OF COBALT

2.1 Chemistry of Cobalt

The predominating oxidation state of Co in aqueous solution is Co(II), but there are some complexing agents, such as ammonia, that will stabilize the Co(III) (BAI 73). Co(II) undergoes hydrolysis - above pH 7.5 this becomes important, and at pH 8.5 a practically total precipitation occurs for trace concentrations in groundwater (RAN 78). The Co(II) hydroxide is amphoteric and dissolves in alkali yielding $Co(OH)_4^{2-}$.

2.2. Parameters of importance for sorption

2.2.1 pH

Meyer et al (MEY 79:3) have measured K_d vs pH ($4 \leq pH \leq 6.5$) for Na-montmorillonite in 1 M NaCl solution. K_d (called D by the authors) increases with increasing pH to a much higher degree than for Sr. An increase in K_d by two orders of magnitude between pH 4 and 6.5 is observed (from $1 \cdot 10^{-3}$ to $1 \cdot 10^{-1}$ m^3/kg).

Means et al (MEA 78) have measured the sorption of ^{60}Co on Conasauga shale at pH 6.7 and 12.0 and obtained the K_d values 70 and 1.2 m^3/kg respectively, i.e. a decrease with increasing pH.

Saas (SAA 79) has shown that the sorption of ^{60}Co on river sediments increases rapidly between pH 6 and 8 and desorption decreases in the same pH range. This is attributed to "the nature of the ion", in this case a formation of sparingly soluble hydroxides.

McKinley et al (MCK 81:2) have measured K_d for sand from Drigg, Cumbria. At pH 4 K_d = 0.0004, at pH 4.5 K_d = 0.0012 and at pH 5 K_d = 0.0016 m^3/kg.

Straub (STR 53) has studied Co-sorption on a clay and found 12% sorption at pH 2.25 and 99.76% at pH 6.40.

Doshi et al (DOS 77) have found a rapid increase of the distribution coefficient (K_d from 1.5 to 4 m^3/kg) between pH 6 and 8.

2.2.2 Nuclide concentration

Meyer et al (MEY 79:2) have studied the distribution coefficient vs loading on alumina. At loadings $\leq 10^{-4}$ moles/kg K_d (called D by the authors) is fairly constant. At higher loadings K_d is decreasing and a total capacity of about $5\text{-}8 \cdot 10^{-2}$ moles/kg is estimated for Co on alumina.

Rançon (RAN 78) has measured the distribution coefficient for ^{60}Co on an altered shale from La Hague. The K_d is fairly constant up to about 10^{-3} M Co.

McKinley et al (MCK 81) have measured the sorption isotherm for Co on a glacial sand from Drigg, Cumbria, UK and on a granitic fracture infill material from Camborne, Cornwall, UK. Experessed as Freundlich isotherms the results are:

Drigg sand, 25°C		$C_R = 39.8\ (C_W)^{0.49}$	(correl. coeff. 0.75)
Camborne fracture infill	25°C	$C_R = 0.77\ (C_W)^{0.33}$	(correl. coeff. 0.75)
- " -	5°C	$C_R = 0.42\ (C_W)^{0.32}$	(correl. coeff. 0.66)

C_R = conc on solid g/g C_W = conc in liquid g/ml
(Note The correlation coefficients ≠ 1.0 indicates quite poor fit to the isotherm.)

2.2.3 Ionic strength

Meyer et al (MEY 79:6) have tried to determine K_d (called D in the paper) by axial filtration through a clay cylinder in 3.6 M and 5 M NaCl solution, but the sorption on Na-montmorillonite was too low to be determined. In 1 M NaCl, the sorption was measurable.

Saas (SAA 79) has compiled data on the sorption of Co on river sediments in groundwater and seawater. A decrease in sorption with ionic strength was observed, since about 60% of the Co was adsorbed in groundwater and only 20% in seawater under similar conditions.

2.2.4 Competing ions and complexation

Meyer et al (MEY 79:3) have studied the Co-sorption on montmorillonites vs $[Na^+]$ and $[Ca^{2+}]$. For pure ion exchange processes, the slopes of K_d (=D) vs $[Na^+]$ and $[Ca^{2+}]$ would be -2 and -1, respectively, but they were found to be steeper. This might be explained by chloride complexation.

Rançon (RAN 78) has studied the influence of $[Na^+]$, $[Ca^{2+}]$ and $[Fe^{3+}]$ on the sorption of Co on an altered shale from la Hague. At $[Fe^{3+}]$ > 10 mg/l the sorption of trace Co is very low. For the concentrations of $[Na^+]$ and $[Ca^{2+}]$ that may be found in groundwater the influence is neglible.
(Note in Swedish groundwaters $[Ca^{2+}]$ up to 100 mg/l have been found (KBS 98). In this case the Co sorption might be influenced).

Barney (BAR 78:2) has tried to correlate the distribution coefficient for some radionuclides to concentrations of some competing and complex forming ions. For Co-sorption the following expression was derived:
$\ln K_d = 6.316 + 4000\,[K^+]^2 - 0.001\,[\text{humic acid}]^2 - 1000\,[NO_2^-]\,[K^+]$.
$[Na^+]$ and $[Ca^{2+}]$ were found to be of less importance. Routson et al (ROU 81) have determined an equation for Ca^{2+} dependence in Stanford soil:
$\ln K_d = 6.74 - 0.932\,X_3 + 0.591\,X_3^2$
where $X_3 = f(\ln\,[Ca^{2+}])$

Means et al (MEA 78) have studied the influence of the complex forming agent EDTA on Co-sorption on Conasauga shale. In the presence of 10^{-5} M EDTA, K_d was decreased form 70 to 0.0029 m^3/kg at pH 6.7 and from 1.2 to 0.0008 m^3/kg at pH 12. The autors conclude that EDTA seems to be responsible for the migration observed in waste burial trenches in Oak Ridge but there seems to be some transport due to humic acid complexation as well. Means (ONW 80:2) has also investigated the influence of oxalic acid on Co-sorption on kaolinite. Adsorption is enhanced by oxalic acid.

Saas (SAA 73) has shown that in calcarous soils, Co is complexed by fulvic acids, while in podzolic soil both humic and fulvic acids act as strong Co-complexing agents.

2.2.5 Ion exchange capacity

Miettinen et al (MIE 81) have tried to correlate the distribution coefficents for Co on a heavy clay, a silty clay, a sandy till and a fracture filling with the cation exchange capacity of these materials, but no correlation was found.

2.2.6 Temperature

Erdal, Wolfsberg et al (ERD 77), (WOL 77), (WOL 78) have measured the distribution coeffcient for Co on alluvium from the Nevada Test Site at room temperature and at $70^{0}C$. K_d (room temp.) = 0.0133 ± 0.012 m^3/kg and K_d ($70^{0}C$) = 0.0180 ± 0.0015 m^3/kg were obtained.

Means (ONW 80:2) has measured K_d vs temperature for Co on kaolinite. Very small changes with temperature were obtained from 25 to $150^{0}C$ in the absence of oxalic acid.

McKinley et al (MCK 81:2) have investigated the sorption of Co on granite at $5^{0}C$ and $25^{0}C$. No significant difference between the distribution coefficients at these temperatures was observed.

2.2.7 Contact time

Erdal, Wolfsberg et al (ERD 77), (WOL 77), (WOL 78) have studied the sorption of Co on alluvium from the Nevada Test Site and on bentonite at contact times up to 144.6 days. Equilibrium seems to be attained for alluvium in about two weeks. The authors have not commented the increase in distribution coefficient with time for bentonite but there seems to be a continuous increase.

McKinley et al (MCK 81:2) have found that Co-sorption reaches an equilibrium after 5-10 days on Alnabreac granite and after 13 days on Drigg Sand.

A fast sorption on MnO_2 was observed by Momma et al (MOM 67) - after 10 hours the sorption was complete.

Straub et al (STR 53) observed a complete sorption on clay after 30 min.

Sumrall et al (SUM 68) concluded that sorption on Zilpha and Yazoo clay occured in less than 90 min.

2.3 Distribution Coefficients

Dlouhy (DLO 81) reports distribution coefficients for Co on sand measured by batch technique (K_d = 0.040 m^3/kg) column technique (K_d = 0.012 m^3/kg) and thin layer chromatography (K_d = 0.011 m^3/kg).

2.3.1 Rocks

Erdal, Wolfsberg et al (ERD 77), (WOL 77), (WOL 78), (WOL 78:2) have studied Co-sorption on alluvium from the Nevada Test site taken 200-300 m below the ground. A geologic characterization of the material has been performed (cf ref WOL 78). In this it may be noted : the alluvium consists of mineral and tuff fragments in an iron-oxide-rich fine-grained matrix. Major constituents are quartz, plagioclase, alkali feldspar, calcite, glass shards, and opaques. Minor constituents are amphibole, biotite, orthopyroxene, clinopyroxene. The surface area was 6-20 m^2/s depending on the grain size. The groundwater used was analyzed (ref WOL 78) and among others found to have a $[HCO_3^-]$ of 170 mg/l and pH 8.5. Wolfsberg (WOL 78:2) recommends a distribution coefficient of 9.0 m^3/kg for sorption and 21.0 m^3/kg for desorption.

Means et al (MEA 78) report K_d-values of 70.0 m^3/kg at pH 6.7 and 1.2 m^3/kg at pH 12 for Conasauga Shale.

Rançon (RAN 78) has obtained a K_d of 0.13-0.14 m^3/kg for an altered shale at trace concentrations of Co.

McKinley et al (MCK 81:2) report distribution coefficients for Altnabreac granite, containing feldspar, quartz, chlorite, mica, kaolin group minerals and amphibole group minerals. The Altrabreac groundwater was analyzed, and found to have a low carbonate content ($[HCO_3^-]$ = 13 mg/l). A quartz sand (Drigg sand) has also been studied. This contains about 95% quartz and minor amounts of feldspar. The remaining 5% was found to be mica, chlorite, kaolinite, and montmorillonite. Water equilibrated with this sand has very low carbonate contact ($[CO_3^{2-}]$ $\leq$ 4 mg/l). K_d for Altnabreac granite at $5^{o}C$, 71 d contact time and 0.025 ppm Co ($4.2 \cdot 10^{-7}$ M) initial concentration was 0.165 m^3/kg. For Drigg sand, pH 5.0, $25^{o}C$, $[Co]$ $8.4 \cdot 10^{-6}$ M K_d was 0.0016 m^3/kg.

Kohtola et al (KOH 79) and Miettinen et al (MIE 81) have measured K_d for Co on sand in groundwater and report K_d = 0.001 m^3/kg. For four different sandy tills the investigators obtained K_d = 0.160, 0.140, 0.400 and 0.880 m^3/kg respectively ($[Co]$ = 10^{-12} - 10^{-6} M, t = 7 d) and for two fracture fillings at the same concentrations and contact times K_d = 9.7 m^3/kg and 2.9 m^3/kg.

2.3.2 Pure minerals

Meyer et al (MEY 79:2), (MEY 79:3), (MEY 79:6) have determined distribution coefficients for Co on some different montmorillonites and alumina (Ficher chromatographic) At low $[Co]$ and low ionic strength, and pH 5, K_d approaches 1.0 m^3/kg for Na-montmorillonite and 0.1 m^3/kg for the Ca-form. A synthetic Na-montmorillonite has a K_d = 4.5 m^3/kg at pH 6.5 and 1.0 M NaCl solution, while a Texas Ca-montmorillonite under the same conditions has K_d = 0.010 m^3/kg.

For alumina, trace concentrations, 0.2 M NaCl solution and pH 6.8 a K_d of 1.0 m^3/kg was obtained.

Means et al (MEA 78:2) have found that a large fraction of the ^{60}Co adsorbed on Oak Ridge soil is found on the manganese oxide. This may be attributed to the large surface area of Mn(IV) minerals ($\leq$ 350 m^2/g) and to the low point of zero charge (ZPC) of Mn(IV) oxides (pH = 1.5-2.0), and to the high cation exchange capacity (CEC = 1.5 eqv/100 g for MnO_2 at pH 8.3).

Means has also studied Co-sorption on kaolinite (ONW 80:2), where K_d at 25^oC, $[Co] = 5 \cdot 10^{-4}$ M, pH 5-6 is 0.2-0.3 m^3/kg.

Skytte-Jensen (SKY 80) reports Danish measurements for Co-sorption on some clay minerals:

Kaolinite	$K_d = 0.003 - 0.32\ m^3/kg$
Illite	$K_d = 1.0 - 10\ m^3/kg$
Montmorillonite	$K_d = 0.1 - 3.2\ m^3/kg$

Doshi et al (DOS 73) have studied removal of Co from seawater by coprecipitation and sorption with MnO_2. This leads to a complete removal. Sorption on ferric hydroxide removes about 83%.
Doshi et al (DOS 77) have also studied sorption on hydrous titanium dioxide. At 400 mg TiO_2/l, 50% of the cobalt is sorbed at pH 7.6.

Momma et al (MOM 67) have measured sorption on different types of MnO_2. The sorption ranged between 28 and 99%.

2.3.3 Soils, clays, sediments etc.

Routson et al (ROU 81) gives an equation for Co-sorption vs $[Ca^{2+}]$. At 0.1 M Ca^{2+} $K_d = 0.85\ m^3/kg$.

Kohtola et al (KOH 79) have studied Co-sorption on a number of clays and soils and obtained K_d-values between 0.25 m^3/kg and 2.5 m^3/kg. Miettinen et al (MIE 81) have obtained K_d-values between 0.410 m^3/kg and 14 m^3/kg for clays and between 0.14 and 0.88 m^3/kg for sandy tills. ($[Co^{2+}] = 10^{-12} - 10^{-6}$ M).

Dlouhy (DLO 81) reports K_d-values of 0.003 - 2.0 m^3/kg for soils. No further data for these measurements are given in the paper.

Sorathesen et al (SOR 60) have studied Co-sorption on some pure clay minerals (illite, kaolinite, montmorillonite and vermiculite) and compared this to sorption on Clinch River sediments. The sorption on the sediments was higher than accounted for only by the mineral composition, and some organic matter interaction was suspected.

Straub et al (STR 53) have measured sorption on 50000 ppm clay in tap water. The sorption vs pH was:

pH	2.75	4.05	4.40	4.45	4.80	5.25	5.50
$K_d(m^3/kg)$	0.0027	0.0182	0.028	0.039	0.095	0.40	0.73

pH	5.80	6.40
K_d	1.94	8.31

Sumrall et al (SUM 68) have measured about 80% sorption on Yazoo and Zilpha Clays.

2.4 Suggested sorption mechanisms

There are not many conclusions drawn on possible sorption mechanisms for Co in the cited literature. Some observations are however made:

- o no correlation between K_d and CEC is obtained for soils (MIE 81)
- o for montmorillonite, the expected solpes of K_d vs $[Na^+]$ and $[Ca^{2+}]$ for an ion exchange process are not obtained (MEY 79:3)
- o Manganese oxide, having a large surface area has a high Co-sorption (MEA 78:2)

Routson et al (ROU 81) suggests that Co-sorption may be controlled in part by precipitation like processes.

3. SORPTION OF NICKEL

Very few studies of sorption of trace amounts of Ni have been reported. Some measurements on rocks (shale and sand) and on clay (montmorillonite) have been reported from Oak Ridge National Laboratory.

3.1 Chemistry of nickel.

The predominating oxidation state of nickel in aqueous solution is + 2 and Ni(II) compounds show many similarities with Co(II)-compounds. At pH below 6.7 Ni is highly soluble in water. Many organic Ni-complexes are water soluble and Ni-EDTA, -DTPA, -HEDTA,

-gluconate, and tartrate-complexes do not precipitate at pH above 6.7.
For oxalic acid at I=0.1, log K_1 = 5.3, for fulvic acid log K_1 = 4.13 - 4.16, and for EDTA log K_1 = 17.5. (WIG 78:2), (BAI 73)

3.2 Parameters of importance for sorption

3.2.1 pH

Meyer et al (MEY 79:6) have measured the distribution coefficient for Ni on Na-montmorillonite with and without acetate buffer at some different pH-values. Without buffer K_d was constant between pH 4.3 and 6.1 (K_d = 0.0044 m^3/kg). With buffer K_d increased from 0.0021 m^3/kg at pH 4.7 to 0.0036 m^3/kg at pH 6.3.

3.2.2 Ionic strength

A decrease in distribution coefficient with increasing ionic strength was found by Meyer et al (MEY 79:3).

Francis et al (FRA 77) have studies K_d for Muscatine silt loam at 0.01 and 0.10 meq/ml $[Ca^+]$ by liquid chromatography and obtained K_d > 0.015 m^3/kg at both concentrations. For Fuquay sand at 0.10 meq/ml $[Ca]$, K_d = 0.0085 m^3/kg and at 0.01 meq/ml K_d = 0.0054 m^3/kg. In batch measurements K_d at 0.1 meq/ml $[Ca]$ was 0.27 m^3/kg for Muscatine silt loam.

3.2.3 Contact time

Francis et al (FRA 77) have found that the sorption of Ni on Muscatine Silt Loam is fast (some hours for complete sorption).

3.3 Distribution coeffecients

3.3.1 Rocks

Francis et al (FRA 77), (FRA 78) have determined distribution coefficients by channel liquid chromatography (CLC) and by batch measurements.

Solid	K_d m^3/kg	
	CLC	Batch
Muscatine Silt Loam	>0.015	0.270
Fuquay Sand	0.00085-0.0054	-
Conasauga Shale	0.0053-0.0086	0.0065

3.3.2 Pure minerals

Meyer et al (MEY 79:3) have determined distribution coeffecients for Na-montmorillonite by axial filtration and obtained K_d values between 0.014 m^3/kg (pH 6.8, low ionic strength) and 0.0001 m^3/kg (pH 4.7, high ionic strength). The Ni-concentration was in all cases about 10^{-4}M.

3.4 Suggested sorption mechanisms

No sorption mechanisms for Ni are suggested in the cited literature.

4. SORPTION OF STRONTIUM

4.1 Chemistry of strontium

In aqueous solution Sr exists only as Sr(II)(BAI 73). Of the anions existing in Swedish groundwaters, CO_3^{2-} and OH^- would dominate the complex formation with Sr. A summary of complex formation constants is given below.

Complex formation constants for Sr (SIL 64):

$SrCO_3(s) \rightleftharpoons Sr^{2+} + CO_3^{2-}$	$\log K_s = -9.03$
$Sr^{2+} + CO_3^{2-} \rightleftharpoons SrCO_3$	$\log K = 3^*$
$H^+ + CO_3^{2-} \rightleftharpoons HCO_3^-$	$\log K_1 = 10.33$
$H^+ + HCO_3^- \rightleftharpoons H_2CO_3$	$\log K_2 = 6.35$
$Sr(OH)_2(s) \rightleftharpoons Sr^{2+} + 2OH^-$	$\log K_s = -3.2^*$
$Sr^{2+} + OH^- \rightleftharpoons SrOH^+$	$\log K_1 = 0.8$

* Values extrapolated from known constants of the alkaline earth metals.

At trace amounts of Sr (< 10^{-7}M) the amount of $SrCO_3$ in a groundwater with 123 ppm $[HCO_3]_{tot}$ will exceed 1% at pH above 8. In this environment hydrolysis is neglible.

4.2 Parameters of importance for sorption

4.2.1 pH

Erdal et al (ERD 78:2) concludes that pH does not seem to have any major effect on Sr-sorption on granite.

Kenna et al (KEN 80) at Sandia Labs have studied the distribution coefficient vs pH for subseabed smcetite sediments between pH 2 and 6.At pH 2 - 4 K_d is constant about 0.030 m^3/kg, and from pH 4 to 6 it increases to 0.10 m^3/kg and further to 0.20 m^3/kg at pH 10.

Brockett et al (BRO 53) have studied the pH dependence of Sr sorption on clays. An increase with pH was found, especially between pH 9 and 10.

Meyer et al (MEY 79), (MEY 79:4), (RAF 81) at Oak Ridge have found that the distribution coefficient for Sr on alumina is increasing significantly with pH for some minerals. Among the clays montmorillonite did not show any change in sorption between pH 5 and 7, while a slight increase was observed for illite and a pronounced pH dependence was found for kaolinite. The same investigators (MEY 80:4) have studied sorption on SiO_2 and silica gel and found a 10-fold increase in distribution coefficient between pH 6 and 9.

Burkholder et al (BUR 79) report a linear regression correlation of K_d vs a number of parameters. Two equations are presented, derived with and without time dependence. Without time dependence and with all other parameters constant, ln K_d is proportional to 1.1 pH (K_d in ml/g) and with time dependence ln K_d is proportional to 0.49 pH. These equations are obtained from a series of laboratory measurements and seems to indicate that the pH dependence varies with time.

Rai et al (RAI 78:3) conclude in a literature survey that Sr and water migrates with the same rate in soil at pH 3.

McHenry et al (MCH 56) have studied Sr-sorption on soils and found a strong pH-dependence. The Sr removal from solution increases from 10% at pH 4 to about 95% at pH 10.

Baudin (BAU 81) has studied sorption of Sr on illite and attapulgite at pH 5.7 and 9 and found no or little pH-dependence of sorption, except for attapulgite, where a higher (2-3 times) sorption at pH 9 compared to pH 5 and 7 is obtained.

Rançon (RAN 78) has studied Sr-sorption on clay and found a very large pH-dependence - an increase in K_d by two orders of magnitude from pH 4 to 10. Rochon et al (ROC 79) have measured Sr-sorption on quartz, illite and vermiculite over a wide pH-range (pH 1-13). A maximum in K_d was found at pH 9-11, followed by a slight decrease at higher pH.

Saas (SAA 79) has studied sorption on an altered shale where K_d increased from 0.002 to 0.500 m^3/kg when pH increased form 1 to 10.

Kepak (KEP 79) has studied Sr-sorption on ferric, ferrous, aluminum and chromic oxides. The pH-dependence indicates that the sorption involves exchange with H^+ on the surface. In alkaline environment, however, there is a decrease in sorption that cannot be explained by formation of less sorbing species (as $SrOH^+$) but some more complicated sorption mechanism must occur.

4.2.2 Nuclide concentration

Sorption isotherms for Climax Stock granite, Eleana argillite and Yucca Mountain tuff have been measured for Sr at 10^{-3}-10^{-6} M initial concentrations (CRO 80), (ERD 80), (WOL 80), (DAN 81:2). The values have been fitted to a Freundlich isotherm ($q = kc^n$) and the exponent n was found to be 0.99-1.11 for granite, 0.88-0.98 for argillite and 0.73-0.87 for tuff, i.e. the distribution coefficient for granite and argillite is close to concentration independent, while a decrease with increasing concentration is obtained for tuff.

Erickson et al (ERI 79) at Sandia have measured the distribution coefficient on abyssal red clay in 0.68N NaCl. At $1 \cdot 10^{-4}$ M Sr K_d = 0.071 m^3/kg and at $4 \cdot 10^{-5}$ M K_d = 0.128 m^3/kg.

Lynch et al (LYN 79) at Sandia have measured Sr sorption on argillite in syntetic groundwater containing 1 ppm Sr carrier and in deionized water with Sr in trace concentrations. The sorption in deionized water was one order of magnitude larger than in the groundwater.

Meyer et al (MEY 78), (RAF 81) have studied Sr sorption in NaCl-solution on montmorillonite kaolinite, attapulgite, and illite and found that measurements at trace concentration are consistent with measurements at higher concentrations, i.e. linear sorption isotherms up to the capacity of the mineral.

Seitz et al (SEI 79) have studied Sr sorption on oolitic limestone in synthetic groundwater at concentrations between 10^{-6} and 10^{-11} M and found no pronounced concentration dependence. The sorption was, however, low at all concentrations.

Relyea et al (REL 79) concludes from a study of Sr-sorption on limestone that K_d seems to be more dependent on final pH than on nuclide concentration.

Isotherm measurements on unconsolidated sediments from Oakwood Salt Dome, Texas, USA (mainly quartz, feldspar and kaolinite) (ONW 80:2) shows that K_d (Sr) is fairly constant at concentrations below 10^{-7} M.

McKinley (MCK 81) has measured sorption isotherms on a rock fracture infill material and obtained Freundlich isotherms as follows:
25^0C: $C_R = 380C_w^{1.08}$ correl. coeff. 0.96
5^0C: $C_R = 195C_w^{1.05}$ correl. coeff. 0.99
(C_R = conc on solid, mol/g, C_w = conc in water mol/ml). The exponents are very close to 1, i.e. the sorption isotherm is almost linear

Rançon (RAN 78) has found a linear sorption isotherm of alluvium from la Durance over a wide range of concentrations.

Bo (BO 79) has found a linear sorption isotherm for Sr on kaolinite.

Heremans et al (HER 76) have measured Sr-sorption on soil from Mol, Belgium at 0.2 mg Sr/l, 10 mg Sr/l and 100 g Sr/l, for a number of soil samples. For 0.2 mg/l $K_d = 0.10\text{-}1.0\ m^3/kg$, for 10 mg/l $K_d = 0.050\text{-}0.200\ m^3/kg$ and for 100 mg/l $K_d = 0.015\text{-}0.060\ m^3/kg$.

Skagius et al (SKA 82:2) have studied sorption on crushed granites of different particle sizes and interpreted the results in terms of a sorption isotherm of Freundlich type ($q = k\ c^n$) (q = mol Sr/kg, c = mol Sr/m^3). For Finnsjö granite $n = 0.93$ and for Stripa granite $n = 1.07$. The konstant k was divided into a volume sorption term (K_V) and a surface sorption (K_A) as follows $K = K_V + 6\ K_A/(\rho_p d_p)$
(ρ_p = particle density (kg/m^3), d_p = particle diametre). The constants were determined to:
Finnsjö: $K_V = 0.79 \cdot 10^{-2}$, $K_A = 0.7 \cdot 10^{-4}$
Stripa: $K_V = 0.13 \cdot 10^{-2}$, $K_A = 0.66 \cdot 10^{-2}$

For the uptake on whole rock surfaces, the same investigators (SKA 82) have determined K to $8.9 \cdot 10^{-4}$ for Stripa granite and $2.7 \cdot 10^{-3}$ for Finnsjö granite, assuming $n = 1$.

4.2.3 Ionic strength

A study of Sr-sorption in two syntectic groundwaters with different ionic strength has been performed in Los Alamos (CRO 79), (ERD 79:5), (ERD 76:6), (ERD 79:9) for tuff, granite, and argillite. The sorption is decreasing with increasing ionic strength.

Dosh et al (DOS 78), (DOS 78:2) have measured the sorption on dolomite in brine and groundwater and found a very low sorption in brine. ($K_d \leq 2 \cdot 10^{-3}\ m^3/kg$) and somewhat higher in groundwater ($K_d = 4.5 \cdot 10^{-3}\ m^3/kg$).

In Oak Ridge, the influence of the ionic strength on Sr-sorption from NaCl-solutions has been studied for a long period of time (MEY 77), (MEY 77:2), (MEY 78), (MEY 78:3), (MEY 79), (MEY 80), (RAF 81). For clay minerals a decrease in distribution coefficient with increasing ionic strength is found, but there is not always the correlation $\log(K_d)$ vs log(ionic strength) as is expected for a pure ionic exchange process. For montmorillonite at pH 7.6, the distribution coefficient decreases from 0.110 m^3/kg at 0.2 M Na^+ to 0.003 m^3/kg at 5.5 M Na^+. In most cases no sorption is observed at 4 M NaCl (RAF 81). For hydrous oxides, the salt dependence is small (MEY 79).

The deviations from ideal ion exchagne behaviour may be attributed to a competition of exchange sites between Na^+ and Sr^+ (MEY 80).

Relyea et al (REL 77) have studied sorption on a number of minerals and found a strong dependence on the salt concentration. The same investigators (REL 79:4) have measured sorption on kaolinite and vermiculite at 5 M and 0.05 M NaCl. For kaolinite K_d changed by a factor of 10, while the difference for vermiculite was a factor of 40.

Serne et al (SER 77) have studied sorption on some rocks and found a slight sorption in distilled water, but no measurable sorption in brine.

Barney (BAR 78:2) gives an equation, obtained by linear regression for Sr-sorption on Hanford soil. When humic acid concentration is constant, the equation is

$$\ln K_d = \text{const} - 97.332 \left[Ca^{2+}\right] - 14.322 \left[Na^+\right] + 507.611 \left[Na^+\right]^2 + 163.449 \left[Ca^{2+}\right] \left[Na^+\right] \qquad (K_d \text{ in ml/g})$$

i.e. there is a dependence on ionic strength indicating that K_d is increasing with ionic strength .

The same author together with Routson et al (ROU 81) gives another equation for Hanford soil: $\ln K_d = -2.13 - 0.864\, f(\left[Na^+\right]) - 1.29\, f(\left[K^+\right]) + 0.207\, f(\left[Ca^{2+}\right]) + 0.558\, f(\left[K^+\right]^2) - 0.108\, f(\left[Ca^{2+}\right]^2) + 0.148\, f(\left[Na^+\right]) \cdot (\left[Ca^{2+}\right])$

The authors conclude from this that $\left[Ca^{2+}\right]$ controls the sorption and that $\left[Na^+\right]$ is of less importance, and thus ion exchange may be the primary sorption mechanism.

Laske (LAS 79) has compared sorption on some clays in distilled water, $Ca(OH)_2$, and $CaSO_4$-solutions. The difference in distribution coefficient is 2-3 orders of magnitude. For tuff a difference of 1-2 orders of magnitude was measured.

Kohtala et al (KOH 79) have studied K_d vs ionic strength for soil and sand. A decrease in K_d by four orders of magnitude when the ionic strength was increased from 0.001 to 1 M(NaCl) was obtained for a clay from Bornholm. For sand the K_d-decrease was only 1-2 orders of magnitude.

4.2.4 Competing ions and complexation

A lower distribution coefficient for Sr on argillite in N_2-atmosphere than in air was observed by the Los Alamos group (DAN 81:2). This may be attributed to the lower carbonate content in water in absence of air.

Hinkebein et al (HIN 77), (HIN 78), (HIN 78:2) at Sandia have studied the irreversible binding of Sr to some rocks. When saturating the rocks with 50 ppm Sr-solution $SrCO_3$ and $SrSO_4$ were formed on the surface of Magenta dolomite, Bell Canyon sandstone and clay bearing halite. Below 10 ppm Sr no precipitation occured. They conclude that the carbonates and sulphates have too high solubility to be of importance for migration of trace Sr.

Spalding (SPA 80) in Oak Ridge has studied the distribution coefficient on Conasauga shale vs organic matter (x), exchangeable acidity (y), and Ca+Mg (z) and obtained the following equation:
$\ln K_d = 3.41 - 0.64x + 0.28y + 0.11z$

Tamura (TAM 62) has shown that Sr-sorption is enhanced by addition of phosphate ions.

Relyea et al (REL 77) at Battelle have obtained high K_d-values for a number of minerals in 0.03 N $NaHCO_3$ compared to 0.03 N $CaCl_2$ solution. Here carbonate complexation may enhance the sorption, while calcium is competing with strontium for sorption sites.The same investigators (REL 79) try to explain the large deviations in measured distribution coefficents for Sr on limestone within the controlled sample program by coprecipitation or occlusion of Sr with monolayers of $CaCO_3$ that are formed by slight termperature changes in the $CaCO_3$ saturated water/rock system.

Ames et al (AME 66) have studied Sr-removal by soils and found a strong correlation between soil calcite content and Sr-sorption.

McFadden (MCF 80) concludes in a literature survey that although Sr-complexes usually are weak, organic matter or phosphate may cause retention of Sr in soils. Detergents also enhances sorption. Microbiological activity may lead to hindered migration. Among competing ions the order of importance has been found to be: Al > Fe > Ba $\geq$ Ca > Mg > H and Cs > K > Na > NH_4.

Barney et al (BAR 78:2) have shown that the content of humic acid in Hanford soil is one of the main parameters influencing sorption.

Ames (AME 60) has studied the Sr-removal by inclusion when calcite is treated with phosphate-rich water (0.005-0.05 M PO_4^{3-}). Some of the CO_3^{2-} is exchanged for PO_4^{3-} and vacant cation sites are formed. If Sr is present it may occupy these sites.

McHenry et al (MCH 56) have shown that Sr-sorption on soils is enhanced by PO_4^{3-}. In absence of PO_4^{3-}, addition of NO_3^- leads to decreased sorption.

Baudin (BAU 81) has studied Sr-sorption on illite and attapulgite with and without addition of humic acid (100 mg/l) to the water. No significant difference was found in the pH-range 5 to 9.

Saas (SAA 73) has shown that some of the added Sr was found in the humic and fulvic acid fraction of water that had been contacted with soil.

Baetsle et al (BAE 62) have measured K_d for soils from different depth at Mol, Belgium and found the highest K_d-values in profiles with high content of organic material.

Dlouhy (DLO 81) has shown that the influence of competing ions for sorption on a sandy soil is Sr > Ca > K > Na.

Skagius et al (SKA 82) have studied simultanous sorption of Sr and Cs on Finnsjö granite surfaces, and obtained a K_d of $3.7 \cdot 10^{-3}$ m^3/kg compared to $2.7 \cdot 10^{-3}$ m^3/kg at sorption of Sr alone.

4.2.5 Ion exchange capacity

Measurements of Sr-sorption on tuff performed in Los Alamos (VIN 80), (WOL 80), (DAN 81:3) shows that the tuffs with high content of zeolitized material (=high CEC) also have the highest distribution coefficients. In column experiments on tuff cores the shape of the elution peak indicates that ion exchange equilibria can account for the sorption.

Tamura (TAM 62) concludes that CEC is of great importance for the sorption.

Burkholder et al (BUR 79) reports a CEC dependence for sorption obtained from linear regression studies of Sr-sorption. When all other parameters are constant the resulting equation is $\ln K_d = \text{const}_1 + 0.49$ CEC with no time dependence or $\ln K_d = \text{const}_2 + 0.60$ CEC with time dependence included (K_d in ml/g).

Miettinen et al (MIE 81) have found correlations between both CEC and exchangeable Ca in soils and K_d for Sr.

4.2.6 Temperature

Sorption of Sr on alluvium, argillite, and granite has been studied in Los Alamos (ERD 77), (WOL 78), (ERD 78:2), (ERD 79:2), (ERD 79:8) at 22^oC and 70^oC. In all cases an increase of the distribution coefficient by about a factor of 2 was observed.

Kenna (KEN 80) at Sandia has studied the temperature dependence of Sr-sorption on subseabed smectite sediments. At 20^oC $K_d = 0.130$ m^3/kg and at 60^oC $K_d = 0.170$ m^3/kg.

Meyer et al (MEY 79) have studied Sr-sorption on illite and montmorillonite at room temperature and at 80^oC. For illite an increase in K_d with about a factor of 2 was observed, while for montmorillonite no increase and in some cases even a decrease was observed.

Heremans et al (HER 79) have measured sorption on Boom clay at different temperatures: At 100^{0}C, K_d = 0.527 m^3/kg, 300^{0}C, K_d = 0.208 m^3/kg and at 500^{0}C K_d = 0.101 m^3/kg.

4.2.7 Contact time

The time dependence of Sr-sorption has been studied in Los Alamos (ERD 80:2), (WOL 77), (WOL 78) for argillite, alluvium and bentonite. For argillite under oxic conditions a small increase in distribution coefficient with time was observed, while for alluvium the equilibrium was reached quite fast (less than 3 days). For argillite in absence of air the sorption seemed to be very little dependent on time.

Kenna (KEN 80) concludes that for sorption on subseabed smectite equilibrium is reached within four days.

Dosh et al (DOS 79:2) have shown that K_d for argillite is increasing with time.

Beall et al (BEA 78) at Oak Ridge have shown that sorpiton of Sr on kaolinite and montmorillonite is rapid (minutes) while attapulgite has much slower kinetics (days).

Burkholder et al (BUR 79) reports a time dependent equation for Sr sorption, obtained by linear regression. With all other parameters constant, the equation is lnKd = const - 0.24t.

Rançon (RAN 78) points out that sorption equilibrium is reached within some hours for Sr on clays.

Skagius et al (SKA 82:2) have tried to interprete the sorption kinetics for granite as diffusion into the rock matrix. This was readily done for large particle sizes (0.1-5 mm have been studied) while the values for small particles did not fit into this interpretation.

4.2.8 Redox potential (Eh)

The Los Alamos group has studied the sorption of Sr in presence of air and under anoxic conditions (N_2, < 0.2 ppm O_2, < 20 ppm, CO_2) for tuff, argillite and granite (Climax Stock) (CRO 79), (ERD 79:5), (ERD 79:9), (CRO 80:2), (ERD 80:2), (ERD 80:3), (VIN 80), (VIN 80:2), (DAN 81:2), (ERD 81:2). For argillite the distribution coefficient was about a factor of 5 smaller under anoxic than under oxic conditions. For tuff there is a decrease of the distribution coefficient by a factor of 2 when changing to anoxic conditions. For granite no significant difference in sorption was found.

Eh is included in the linear regression equation reported by Burkholder et al (BUR 79). The Eh-term in the equation has different sign with and without time dependence - without time: $\ln K_d = \text{const}_1 + 0.60\ \text{Eh}$ and with time: $\ln K_d = \text{const}_2 - 0.12\ \text{Eh}$ - when all other parameters are constant (K_d in ml/g).

Relyea et al (REL 79:5) have measured Sr-sorption on 12 different minerals under oxic (air) and anoxic (< 2000 ppm O_2) atmosphere. They found differences in dissolution of minerals under these conditions - Si and Al were lower in the presence of air. This may explain why Sr, that has only the valence state +II in aqueous solutions, is sorbed somewhat differently with and without air. In general, K_d(Sr) was lower under anoxic conditions, but in 5.13 M brine, K_d for illite and vermiculite was increased by a factor of 100 when changing to anoxic conditions. Biotite behaved in a more "normal" way - the sorption decreased by a factor of 2 under anoxic conditions.

4.2.9 Liquid/solid, particle size, surface area

The influence of particle size and solid/liquid ratio in bach sorption has been studied in Los Alamos (ERD 77), (WOL 78), (ERD 80:2), (VIN 80), (DAN 81:2), (DAN 81:4) for granite, tuff, and alluvium. An increased sorption with increasing availible surface (i.e. decreasing particle size or decreased liquid/solid ratio) seems to exist in some cases, but for some materials, as tuff, a mineral fractionation when sieving seems to influence the results.

The linear regression equation reported by Burkholder et al (BUR 79) involves a minor surface area (SA) dependence - $\ln K_d = const_1 + 0.07 [SA]$ or $\ln K_d = const_2 - 0.05 [SA]$ (without and with time dependence, K_d ml/g).

Rançon (RAN 78) has varied the liquid/solid ratio from 100 ml/g to 1 ml/g for clay/groundwater and found no significant change in distribution coefficent.

4.3 Distribution coefficients

Most distribution coefficients reported are measured by batch methods. Some measurements are however performed in crushed rock columns and in fractured rock cores. Comparisons of values obtained by different methods have been done for tuff, argillite and granite (ERD 77), (ERD 79:5), (ERD 79:6), (ERD 79:9), (ERD 80:2), (VIN 80), (VIN 80:2), (WOL 81:3), (DAN 81:3). Most values were a factor 2-3 higher in batch measurements than in column.

Ryan (RYA 81) has compared batch and column measurements for Savannah River soil, and concludes that more reliable information about nuclide movements in soil is obtained by column studies than by batch methods.

Behrens et al (BEH 81:2) have compared bach and column determined K_d for fluviatile and aeolian sand. The agreement was good without any systematic differences.

The question weather distribution coefficients are additive for mixtures of minerals is often raised. At Oak Ridge a study of mixtures of clays and oxides has been performed (MEY 80), (PAL 81). Most mixtures studied showed additive distribution coefficients but for illite + Al_2O_3 lower values were obtained, probably due to saturation of the sorption sites of illite by Al dissolved from Al_2O_3. The same effect is observed for SiO_2 + ZrO_2, where a precipitation of dissolved Si on ZrO_2 might be the explanation.

Komarneni et al (KOM 79) have found additive K_d-values for vermiculite + zeolite or mixtures of two zeolites.

4.3.1 Rocks

From Los Alamos, meaurements of Sr-sorption on granite, tuff, alluvium and argillite have been presented in a number of reports since 1977 (ERD 77), (WOL 77), (ERD 78), (ERD 78:2), (ERD 78:3), (WOL 78), (ERD 79), (ERD 79:2), (ERD 79:3), (ERD 79:5), (ERD 79:6), (ERD 79:7), (ERD 79:8), (ERD 79:9), (WOL 79), (CRO 80:2), (ERD 80), (ERD 80:2), (ERD 80:3), (VIN 80), (VIN 80:2), (WOL 80), (DAN 81:2), (DAN 81:3), (DAN 81:4), (ERD 81:2),(WOL 81:3). Studies of retention of Sr in the ground has also been performed in connection with bomb tests in Nevada (HOF 78) and the Oklo reactor (BRY 76), (COW 77), (COW 78).For granite (Climax Stock) distribution coefficients between 0.004 m^3/kg and 0.030 m^3/kg at 22°C and 0.021-0.151 m^3/kg at 70°C have been measured. Tuffs from different locations (Yucca Mountain, Jackass Flats, and G-tunnel, Nevada Test site) have been studied. The mineralogy varies much within the same location. The distribution coefficients vary between 0.041-25 m^3/kg (Yucca Mountain), 0.035-45 m^3/kg (Jackass Flats), 39-54 m^3/kg (G-tunnel) at 22°C. At 70°C the Jacass Flats samples have distribution coefficients in the range 0.106-22.2 m^3/kg. For alluvium, distribution coefficients of 0.048-0.629 m^3/kg were obtained.

Average distribution coefficient values (in m^3/kg) are given for some solids (ERD 78):

Quartz Monzonite (granite):	0.020 (22°C)	0.040 (70°C)
Argillite :	0.130 (22°C)	0.290 (70°C)
Alluvium :	0.200 (22°C)	
Vitric tuff :	13.00 (22°C)	14.00 (70°C)
Devitrified tuff :	0.060 (22°C)	0.110 (70°C)
Zeolitized tuff :	0.240 (22°C)	1.00 (70°C)

At Sandia, distribution coefficients have been measured for natural as well as for synthetic dolomite, for argillite and hornfels (DOS 78), (DOS 78:2), (DOS 79), (DOS 79:2), (HIN 77). Low distribution coefficients have been obtained for dolomite (< 0.002 m^3/kg in brine, 0.001-0.032 m^3/kg in groundwater). For natural dolomites (Magenta and

Culebra) the distribution coefficient varies with one order of magnitude, indicating that these materials are inhomogenous. Synthetic dolomites give values in the same range.

For argillite, the distribution coefficient in groundwater was 0.040-0.070 m^3/kg and for hornfels 0.020 m^3/kg (LYN 79). Dosh et al (DOS 79:2) reports 0.021-0.060 m^3/kg on Eleana argillite in simulated groundwater. Both these measurements were carried out at high (5-15 ppm) Sr-carrier concentration.

Brockett et al (BRO 53) have measured sorption of fission products on Conasauga shale from a tap water solution and obtained a distribution coefficient of 0.63 m^3/kg.

Francis et al (FRA 77), (FRA 78), (FRA 78:2) have measured distribution coefficients for Muscatine silt loam by channel chromatography and obtained low values, 0.001 - 0.004 m^3/kg, at 0.1 meq/cm^3 Ca^{2+}. At 1 M Ca^{2+} no sorption was observed, but at 10^{-4} M Ca^{2+} K_d = 1 m^3/kg. Meyer et al (MEY 78), (MEY 78:3) have measured K_d for limestone and basalt within the controlled sample program. In groundwater K_d for limestone = 0.005-0.006 m^3/kg and for basalt up to 0.108 m^3/kg at pH 8. In brine K_d for both rocks was about 0.001 m^3/kg.

Borg et al (BOR 76) have compiled K_d-data for rocks measured at some US laboratories . Since these reports were not availible in this study, some data are quoted:

Solid	K_d m^3/kg	Water
Alluvium	0.048-2.45	Groundwater
Tuff, Rainier Mesa, NTS	0.260-4.3	Prepared water
Tuff, NTS	0.400	Groundwater
- " -	4.00	Not given
Carbonate, Yucca Flat	0	Prepared water
Dolomite, Rustler Formation NM	0.005-0.014	- " -
Granodiorite, Climax Stock	0.004-0.023	- " -
Granite, Nevada	0.002	Deep formation water
Basalt, NTS	0.016-0.135	Prepared water
Pillow Basalt, Amchitka	0.220	- " -
Basalt, Amchitka	1.220	- " -
- " -	0.001	Sea water
Shaley siltstone, NM	0.008	Well water
Sandstone, NM	0.001	- " -

MacLean et al (MAC 78) at Lawrence Livermore Laboratories have determined distribution coefficients for some rocks:

Solid/aqueous phase	K_d m^3/kg
Limestone + groundwater	0.003
- " - + brine	0.001
Basalt + groundwater	0.045
- " - + brine	0
Dolomite + groundwater	0.001
Climax Stock Granite + groundwater	0.059
Westerly granite + - " -	0.007
Conasauga shale + - " -	0.040
Tuff + - " -	0.029
Green argillaceous shale + groundwater	0.060
Climax shear zone^x + groundwater	0.175-0.608

^{x}mostly montmorillonite + calcite

Coles et al (COL 79), also at LLL, have studied sorption on 15 different sandstones in brine. No sorption of Sr was observed. In column studies of tuff and sandstone, Sr is retained by tuff but not by sandstone.

Weed et al (WEE 81) at LLL have found a retention factor of 1.2 for Sr in sandstone in dynamic experiments.

Seitz et al (SEI 78) at Argonne, have tried to measure retention of Sr in columns of crushed oolitic limestone. Very low retention was obtained. The migration was about 10 times faster than would be expected from batch measurements. The same investigators (SEI 79) have performed batch measurements for Sr and oolitic limestone in groundwater at Sr-concentrations between 10^{-6} and 10^{-11} M. K_d was low - 0.004-0.007 m^3/kg.

Burkholder et al (BUR 79) at Battelle reports the following K_d-value: Basalt,(0.003 N $NaHCO_3$, pH 7.9), 0.150 m^3/kg

Relyea et al (REL 77) report the following distribution coefficients:

Solid	5.13 N NaCl	0.03 N $CaCl_2$	0.03 N $NaHCO_3$
Shale	0.003 m^3/kg	0.009 m^3/kg	0.4 m^3/kg
Granite	0.001-0.006 m^3/kg	0.004 m^3/kg	0.03 m^3/kg
Limestone	0.001-0.023 m^3/kg	0.005 m^3/kg	10.0 m^3/kg
Basalt	0.001-0.012 m^3/kg	0.003 m^3/kg	0.1 m^3/kg

Serne et al (ONW 80:2) reports for unconsolidated sediments (Oakwood Salt dome, Texas, USA, mainly quartz, feldspar, and kaolinite), a distribution coefficient of 0.140-0.180 m^3/kg

Serne et al (SER 77) also reports distribution coefficients in distilled water for dolomite (0.006-0.012 m^3/kg), limestone (0.010 m^3/kg) and sandstone (0.019-0.038 m^3/kg).

The results of an interlaboratory comparison of distribution coefficients for limestone and basalt (SER 79) is given below: (measurements where pH is not given are omitted here)

Lime stone		Basalt	
pH	K_d m^3/kg	pH	K_d m^3/kg
7.6	0.002	7.4	0.041
7.9	0.015	8.1	0.068
8.0	0.005	8.4	0.081
8.3	0.006	8.4	0.089
8.5	0.001	8.7	0.092

Komarneni (KOM 79) has not obtained any Sr-sorption on Conasauga shal from a waste solution at pH 7.

Barney et al (BAR 77) report distribution coefficients for some rocks in synthetic groundwater:
basalt: K_d = 0.151 m^3/kg, granite (Climax Stock): K_d = 0.005 m^3/kg, argillite (Eleana): K_d = 0.191 m^3/kg. Barney et al (BAR 79) also reports distribution coefficients for artificially weathered rocks. The weathering has been performed by continuous leaching in distilled water for six months either in contact with air or in nitrogen atmosphere. The distribution coefficients were: (all values in m^3/kg)

Rock	not weathered	weathered in air	weathered in nitrogen
Basalt	0.108	0.500	0.034
Granite	0.006	0.006	0.004
Argillite	0.108	0.078	0.041

Brookins (BRO 76), (BRO 78) concludes from studies of the Oklo phenomenon that the ^{90}Sr probably has been redistributed and only 0.5-5% is retained.

Pickens et al (PIC 80) report distribution coefficients for material from a Candian aquifer, mainly consisting of feldspar and quartz, to 0.003-0.011 m^3/kg.

In a study at AECL in Whiteshell, (MAY 79), measurements of distribution coefficients for some rocks in groundwater were performed:

Rock	K_d m^3/kg
Granite,	0.003
Gabbro,	0.008
Syenite,	0.015
Anothorsite,	0.025

Calmels et al (CAL 79) have performed in situ experiments in fissured granite and obtained no retention (or the same retention as for Rhodamine WT) but a low recovery of Sr (31%). This may be interpreted as an irreversible sorption of part of the Sr, but only 38% of the Rhodamine is recovered, i.e. the same should be valid for this.

Goblet et al (GOB 81) have found a retention of Sr in field experiments and interpreted the recovery in terms of a slow desorption.

Rançon (RAN 67) has measured K_d for calcareous rocks. The value ranged from 0.005 to 0.014 m^3/kg, depending on particle size.

Kohtala et al (KOH 79) have measured a K_d-value of 0.005 m^3/kg for sand in groundwater. Miettinen et al (MIE 81) obtained K_d = 0.150 and 0.180 m^3/kg for two Finnish fracture fillings at 20°C and Sr-concentration 10^{-6}-10^{-12} M.

Tschurlovits (TSC 79) has measured distribution coefficients for some austrian granites at pH in distilled water. K_d was 0.003-0.015 m^3/kg.

Berak (BER 67) has measured distribution coefficients for a large number of silicate rocks for trace amounts of Sr in synthetic river water. The K_d-values are mostly 0-0.050 m^3/kg, in some cases up to 0.200 m^3/kg. Some examples:
Rhyolites 0-0.056 m^3/kg, quartz porphyrites 0.006-0.018 m^3/kg, diabases 0-0.045 m^3/kg, basalts 0.006-0.187 m^3/kg.

Dlouhy (DLO 67) has measured distribution coefficients for tuff from Casaccia, Italy and obtained values of 0.045-0.075 m^3/kg in batch measurements and 0.040-0.042 m^3/kg in column.

Furnica et al (FUR 73) have obtained distribution coefficients of 0.002-0.010 m^3/kg for Sr on marble.

Inoue (INO 67) reports K_d = 0.071 m^3/kg for sand.

For three different tuffs.Laske (LAS 79) has obtained distribution coefficients of 1, 0.6 and 0.5 m^3/kg respectively.

4.3.2 Pure minerals, oxides

Meyer et al (MEY 79), (MEY 79:2), (MEY 79:4), (MEY 80), (MEY 80) have determined K_d for Sr on alumina in NaCl-solutions (4 M - 0.5 M). At pH 7, K_d = 0.100 m^3/kg. At low ionic strength (0.01 M Na^+) and at pH 7 K_d is 1.0 m^3/kg.

Rimshaw et al (RIM 60) have measured Sr-removal from dilute alkaline waste solutions by about 120 different minerals. A rapid exchange with Ca in Florida Pebble phosphate was observed.

Burkholder et al (BUR 79) reports the following data:

Mineral	Water phase	K_d m^3/kg
Albite	0.03 N $NaHCO_3$, pH 9.4	0.550
Albite	0.3 N $NaHCO_3$, pH 7.7	0.004
Anorthite	0.3 N $NaHCO_3$, pH 9.4	0.960
Anorthite	0.3 N $CaCl_2$, pH 7.7	0.003

Relyea et al (REL 77) report measurements for minerals in three different waters (K_d in m^3/kg):

Mineral	5.14 N NaCl	0.03 N $CaCl_2$	0.03 N $NaHCO_3$
Biotite	0.004-0.006	0.012	0.6
Quartz	0.001	0.001	0.3
Albite	0.004-0.006	0.005	0.5
Anorthite	0.003-0.007	0.003	0.4
Microcline	0.003-0.004	0.003	0.4
Hornblende	0.002-0.003	0.002	0.3
Enstatite	0.002	0.001	0.1
Augite	0.005-0.006	0.004	0.8

Komarneni et al (KOM 79) have studied Sr-sorption on some natural zeolites and oxides from a waste solution at pH 7. The following distribution coefficients were obtained:

Mineral	Origin	K_d m^3/kg
Clinoptilolite	Arizona	0.088
Clinoptilolite	Idaho	0.750
Mordernite	Arizona	0.282
Mordernite	Nevada	0.583
Erionite	Oregon	0.019
Erionite	California	0.025
Chabazite	Arizona	0
Phillipsite	California	0
Ferric oxide (red)		0.079
Ferric oxide (yellow)		3.9
Gibbsite		0

Ames (AME 62:3) has measured Sr-sorption on a number of natural zeolites for 0.01 N Sr^{2+} and obtained the following K_d-values:

Mineral	K_d m^3/kg
Clinoptilolite	0.102
Mordernite	0.038
Chabazite	0.028
Stilbite	0.008
Analcite	0.006
Sodalite	0.004

Rançon (RAN 78) has measured distribution coefficients for calcite, $K_d = 0.030$ m^3/kg, and alumina, $K_d > 10$ m^3/kg.

Rochon et al (ROC 79) have measured a K_d for quartz at pH 7-8 of 0.070-0.090 m^3/kg in a synthetic groundwater.

Doshi et al (DOS 77) have measured a distribution coefficient of 0.1 m^3/kg for hydrous titanium dioxide.

Berak (BER 63) has measured K_d-values for trace amounts of Sr in synthetic river water for a large number of silicate minerals. The values were low and no mineral that selectively sorbs Sr was found.For the differnt silicate groups, the distribution coefficients were in the following range:

Silicate group	K_d m^3/kg	
Tectosilicates	0-0.009	(quartz etc)
Tectosilicates	0-0.210	(zeolites)
Nesosilicates	0-0.035	
Sorosilicates	0.014	
Cyclosilicates	0-0.015	
Inosilicates	0-0.030	
Phyllosilicates	0-0.217	

Furnica et al (FUR 73) have obtained the following distribution coefficients for some minerals:

Mineral	K_d m^3/kg
Montmorillonite	$\geq$ 25
Diatomite	0.089-0.173
Kiselguhr	0.016-0.023
Silicagel	0.010-0.014
Kaolin	0.036-0.055

MacLean et al (MAC 78) at Lawrence Livermore Laboratories report a K_d for Sr on biotite in groundwater of 0.059 m^3/kg.

4.3.3 Soils, clays, sediments

For bentonite, some sorption measurements have been performed in groundwater (ERD 77), (WOL 77). Distribution coefficients between 0.900 and 2.90 m^3/kg were obtained. A recommended value for sorption is 1.7 m^3/kg at room temperature.

Erickson et al (ERI 79) have measured K_d for abyssal red clay in 0.68 N NaCl. K_d at $1 \cdot 10^{-4}$ M Sr was 0.071 m^3/kg and at $4 \cdot 10^{-5}$ M 0.128 m^3/kg.

For smectite sediments, Kenna (KEN 80) has measured distribution coefficients in the range 0.1-0.2 m^3/kg.

Nowak (NOW 80) has not found any appreciable sorption on bentonite and hectorite in brine at $6 \cdot 10^{-5}$ M or $2 \cdot 10^{-4}$ M initial Sr concentration.

Beall et al (BEA 78) at Oak Ridge have measured sorption on attapulgite in brine at pH 5. At 0.25 M Na, K_d for attapulgite was 0.007 m^3/kg.

Brockett et al (BRO 53) have measured the following distribution coefficients:
Kaolinite: 0.43 m^3/kg, halloysite: 0.80 m^3/kg, montmorillonite: 1.25 - 1.46 m^3/kg,montronite 1.25 m^3/kg, metabentonite: 0-60-0.75 m^3/kg, vermiculite: 0.95 m^3/kg.

Meyer et al (MEY 77:2), (MEY 78), (MEY 78:3), (MEY 79), (RAF 81), (SHI 79) have measured distribution coefficients for a number of clays (mostly montmorillonite). At low ionic strength (0.2 M Na) and pH 7.6 K_d = 0.110 m^3/kg for montmorillonite, and at 0.1 M 0.080 m^3/kg. Values between 0.01 and 4.3 m^3/kg have been reported for Na-montmorillorite, depending on ionic strength. There are large differences between samples of different origin. For attapulgite distribution coefficients between 1.9 m^3/kg (0.005 M Na) and 0.006 m^3/kg (0.2 M Na) have been measured. Illite gives K_d between 13.01 m^3/kg (0.005 M Na) and 0.008 (0.2 M Na).

Sorathesn et al (SOR 60) have measured the distribution coefficient for Sr on kaolinite to 4 m^3/kg.

Spalding (SPA 80) has measured distribution coefficients between 0.008 and 0.058 m^3/kg for soil (weatherd Conasauga shale) from Oak Ridge.

Tamura (TAM 62) has measured a K_d of 0.015 m^3/kg for Sr on kaolinite.

Relyea et al (REL 77) reports the following K_d-values (m^3/kg) for some clays:

Clay	5.13 N NaCl	0.03 N $CaCl_2$	0.03 N $NaHCO_3$
Illite	0.001-0.003	0.006-0.008	0.5
Montmorillonite	0.001-0.003	0.028	1.2
Vermiculite	0.002	0.023-0.028	1.0
Kaolinite	0	0.001	0.2

Relyea et al (REL 79:4) have compared distribution coefficients for kaolinite and vermiculite with or without pretreatment of the clay in H_2O_2 at 60°C:

Kaolinite,	(0.05 N NaCl)	untreated	K_d = 0.011 m^3/kg
- " -	- " -	H_2O_2-treated	K_d = 0.015 m^3/kg
Vermiculite,	- " -	untreated	K_d = 0.260 m^3/kg
- " -	- " -	H_2O_2-treated	K_d = 0.194 m^3/kg

i.e. the influence of strongly oxidizing agents on these clays seems to be of minor importance for the sorptive properties.

Brown (BRO 67) has measured the distribution coefficient for Hanford soil in natural water and trace Sr to 0.050 m^3/kg.

Komarneni et al (KOM 79) have measured Sr-sorption on montmorillonite and vermiculite from a waste solution at pH 7 and found no measurable sorption.

Lacey (LAC 54) reports sorption data for clay from Oak Ridge (mainly montmorillonite and kaolinite) in Oak Ridge tap water. The distribution coefficients were in the range 4-5 m^3/kg.

Ryan (RYA 81) reports values for Savannah River soil between 0.001 and 0.010 m^3/kg.

Baudin (BAU 81) reports for illite, pH 7, K_d = 0.140 m^3/kg at Sr concentrations at the ppm-level.

Rançon (RAN 79) report K_d-values for clays:

Clay	pH	K_d m^3/kg
Attapulgite	7.5	0.7
Bauxite	7	0.1
Bentonite	10	4
Clinoptilolite	6.5	1
Illite	7.5	1
Kaolinite	6.5	0.1
Sepiolite	8.5	0.15
Vermiculite	8.5	0.2

Rochon et al (ROC 79) have measured K_d for illite and vermiculite. At pH 7-8, K_d (illite) = 0.2-0.3 m^3/kg and K_d (vermiculite) 0.010 m^3/kg.

Laske (LAS 79) has measured sorption on a number of clays and marls in distilled water as well as in $Ca(OH)_2$ and $CaSO_4$ solutions.

Solid	K_d, m^3/kg		
	dist. w.	$Ca(OH)_2$	$CaSO_4$
"Opalinus" clay[1]	0.190	0.014	0.010
"Molasse" clay A[2]	2	0.009	0.016
"Molasse" clay B[3]	1	0.020	0.030
"Molasse" clay B[4]	0.280	0.004	0.006
"Bänder" clay[5]	0.068	0.014	0.014
"Keuper" clay[6]	0.330	0.017	0.020
"Bolus" clay[7]	0.370	0.025	0.015
Kaolin	0.145	0.012	0.011
Marl "Felsenau"[8]	0.033	0.002	0.008

Main minerals in order of occurence:
1 = illite, kaolinite, chlorite
1 = montmorillonite, illite, kaolinite
3 = montmorillonite, quartz, illite
4 = illite, calcite, chlorite, quartz
5 = calcite, illite, quartz, feldspar
6 = corrensite, illite, quartz
7 = kaolinite, Fe-oxide-hydroxide, quartz
8 = dolomite, illite/muscovite, vermiculite, quartz

Kohtala et al (KOH 79) have measured distribution coefficients for a heavy clay and a muddy clay of 0.174 m^3/kg and 0.245 m^3/kg respectively.

Miettinen et al (MIE 81) have determined distribution coefficients for five different heavy clays (K_d = 0.065, 0.054, 0.009, 0.004 and 0.110 m^3/kg, respectively), one silty clay (K_d = 0.042 m^3/kg) and four sandy tills (K_d = 0.029, 0.013, 0.013 and 0.019 m^3/kg respectively) in groundwater at 20°C and 10^{-6}-10^{-12} M Sr .

Skytte-Jensen (SKY 80) reports Danish K_d-measurements for clays at 10^{-3}-10^{-1} M Na^+ or Ca^{2+}. The distribution coefficients ranged over several orders of magnitude:
Kaolinite K_d = 0.010-0.310 m^3/kg

Illite K_d = 0.004-1.0 m^3/kg
Montmorillonite K_d = 0.010-32 m^3/kg

Carlsson et al (CAR 78) have measured K_d-values for some Swedish soils:

Sand, Brunkebergsåsen, Stockholm[1]	K_d = 0.019 m^3/kg
Clay,- " - [2]	K_d = 0.076 m^3/kg
Silt, Porsön, Luleå[3]	K_d = 0.026 m^3/kg
-"- , - " - [4]	K_d = 0.004 m^3/kg
Sand, - " - [5]	K_d = 0.003 m^3/kg

Main minerals:
1 = quartz, feldspar, mica, amphibol
2 = illite, mica, quartz, feldspar, chlorite
3 = quartz, feldspar, mica, illite, montmorillorite
4 = quartz, mica, feldspar, illite, vermiculite, chlorite, hornblende
5 = quartz, feldspar, hornblende, mica

Baetsle et al (BAE 62) have measured K_d for soil from Mol at pH 4 and obtained values between 0.001 and 0.160 m^3/kg.

Levi et al (LEV 67) have studied diffusion of Sr in vermiculite. In alkaline earth saturated vermiculite the diffusivity was about 10^{-11} m^2/s, but if the vermiculite was saturated with Cs, which is fixed by the clay, the diffusivity was far lower.

Godse et al (GOD 67) have measured K_d for Trombay soil at pH 6.8. In groundwater K_d-values in the range 0.143-0.315 m^3/kg were obtained. In water, leached from the soil, lower K_d-values were obtained (0.029-0.077 m^3/kg).

Dlouhy (DLO 67) has determined the distribution coefficient for Cassaccia soil (from Italy) to 0.006-0.015 m^3/kg (batch) or 0.007-0.008 m^3/kg (column). Dlouhy (DLO 81) has also determined K_d for a sandy soil to 0.013 m^3/kg (batch) or 0.022 m^3/kg (column).

4.4 Suggested sorption mechanisms

The sorption mechanism for Sr seems to be mainly a non-selective ion exchange process, as has been observed by many authors, for instance Daniels et al (DAN 81:3), and no Sr selective mineral was found by Berak (BER 63). Levi (LEV 67) also observes that Sr is not fixed in vermiculite as is Cs.

Linear sorption isotherms for rocks and minerals have been observed (DAN 81:2), (RAN 78), (SKA 82:2).

There are correlations between K_d and CEC and between K_d and exchangeable Ca in soil (MIE 81). Wolfsberg et al (WOL 80) have also found a correlation between degree of zeolitization (i. e. CEC) and K_d for tuff.

A quite good description of the Sr-Na exchange in montmorillonite, attapulgite and kaoline is given by ion-exchange theory (MEY 78), (MEY 79), although K_d vs Na for γ-alumina does not show the expected slope for an ion exchange process (MEY 80).

Complexation gives deviations from the ideal ion exchange behaviour in some cases. Rochon et al (ROC 79) have obtained a maximum in K_d vs pH in synthetic groundwater and Saas (SAA 79) concludes that for schiste in natural water ion exchange is the sorption mechanism at pH 6-7, while for higher pH adsorption/precipitation occurs.

5. SORPTION OF CESIUM

5.1 Chemistry of cesium

Cesium in groundwater solution will exist as Cs^+ over a wide pH and concentration range. None of the common complex forming agents found in groundwater will form strong complexes with Cs (SIL 64).

5.2 Parameters of importance for sorption

5.2.1 pH

According to Erdal (ERD 79:2), pH has no major effect on Cs-sorption on granites from Climax Stock and Nevada Test Site.

Dosh et al (DOS 79:2) have found a slight pH-dependence in the pH-range 5.8 - 8.5 for Cs-sorption on Eleana argillite (distribution coefficient changes from 0.040 to 0.068 m^3/kg).

Kenna (KEN 80) has found a small pH dependence for Cs-sorption on smectite clay from subseabed sediments (mainly illite, clinoptilolite and montmorillonite).

Meyer et al (MEY 79:5) have observed a higher K_d for Na-montmorillonite at pH 8 than at pH 5 and the same authors (MEY 80) have shown that the slope of K_d vs pH for Cs-Na-montmorillonite fits quite well with that predicted by ion-exchange theory. For Cs-sorption on SiO_2 Meyer et al (MEY 80:4) have found essentially no pH-dependence.

Silva et al (SIL) have found a very slight pH-dependence for Cs-sorption on montmorillonite (Belle Fourche Clay).

Burkholder et al (BUR 79) report K_d-values for basalt, anorthite, albite and montmorillonite, and observes that the pH-dependence seems to be less important for Cs than for Sr.

Brown et al (BRO 55) have measured sorption of Cs on Hanford soil and obtained 50% sorption at pH 0, 99% at pH 4 and 93% at pH 12-14.

McHenry (MCH 54) has shown that the sorption on Hanford soil was more sensitive to concentration of H^+ than to concentration of any other ions.

Gilham (GIL 80) has determined distribution coefficients for fifteen different Canadian unconsolidated geologic materials (all clays containing illite, vermiculite, montmorillonite, etc.) and found no dicernable correlation between K_d and pH.

McKinley et al (MCK 81:2) have observed a maximum in Cs-sorption on Altnabraec granite at pH 5. For Drigg sand, these investigators report decreasing K_d with pH: pH 4.5, K_d = 2.5 m^3/kg, pH 6, K_d = 0.43 m^3/kg and pH 6.5, K_d = 0.07 m^3/kg.

Baudin (BAU 81) reports an increase in K_d by a factor of 2 between pH 5 and 9 for illite and attapulgite.

Rançon (RAN 78) reports a maximum in Cs-sorption on two different soils (one schiste and one sand) at a pH around the "natural" pH. This may be due to changes in ionic strength when adding NaOH or HCl in order to change pH.

Saas (SAA 79) reports the same maximum for Cs-sorption on altered schiste as Rançon , and he also has data showing that there is no maximum for Sr-sorption under the same conditions.

Rançon (RAN 79) has determined K_d for a number of clay minerals at pH 5 to 9 in synthetic granitic water. For attapulgite and bauxite a minimum is obtained at pH 6-7 while an increase with increasing pH is obtained for bentonite illite, kaolinite and sepiolite and a slight decrease with pH for clinoptilolite. For vermiculite K_d is approximately constant between pH 5 and 8, but decreases from 9 m^3/kg to 1.5 m^3/kg when pH is increased to 9.

5.2.2 Nuclide Concentration

Erdal et al (ERD 80:2) have found a strong correlation between Cs-concentration and distribution coefficient for Cs on granite and argillite. The same investigators have measured sorption isotherms for granite, argillite, and tuff. The isotherms seem linear, i.e. $q = kc^n$ with $n \approx 1$ (Freundlich isotherm) up to about 10^{-8} M Cs in the liquid. At higher concentrations the exponent is lower. Average exponents up to about 10^{-4} M are: granite 0.73, argillite 0.80 and tuff 0.89 and 1.0 (two different samples).

Dosh et al (DOS 79:2) have studied sorption on Eleana argillite and found an almost constant K_d from 0.0012 to 0.2 ppm Cs. Between 25 ppm and 2000 ppm K_d is decreasing from 0.419 to 0.010 m^3/kg. Dosh et al (DOS 79) have also studied sorption on Culebra dolomite (contains also calcite and quartz) and found one isotherm at $8 \cdot 10^{-11}$ to $8.4 \cdot 10^{-8}$ M - a Freundlich isotherm with the exponent 0.87 - and one at $8 \cdot 10^{-8}$ to $7.5 \cdot 10^{-4}$ M - Freundlich isotherm with the exponent 0.75.

Lynch et al (LYN 79) have found no evident concentration dependence in the 0.01-0.0001 ppm range for argillite and hornfels.

Erickson (ERI 79) has observed a sharp increase in K_d between $3 \cdot 10^{-6}$ M and $3 \cdot 10^{-8}$ M Cs for abyssal red clay. He explains the behaviour of this clay (smectite, iron-rich, contains phillipsite) by two different sorption mechanisms - one with high K_d and low capacity and one with lower K_d but higher capacity.

Jacobs (JAC 62) has shown, that for Cs-sorption on vermiculite, there is a sharp increase in Cs-removal at high concentration. He attributes this to a reduction of the (001) lattice spacing to 10Å, causing a physical entrapment of interlayer exchangeable cations.

Tamura (TAM 60) has compared the behaviour of illite, vermiculite and montmorillonite. At low Cs concentration illite has a very high affinity for Cs, and vermiculite is also very effective for Cs-removal, but at higher concentration montmorillonite surpasses these.

Serne et al (ONW 80:2) have found a strong concentration dependence for sorption on kaolinite-containing sediments from Oakwood salt dome, Texas. In the concentration range from 10^{-3} to 10^{-6} M a Freundlich isotherm with exponent 0.32 may be calculated from their data.

Seitz et al (SEI 79) presents data for Senteniel Gap basalt between 10^{-4} and 10^{-9} M. Here a Freundlich isotherm exponent of 0.2 may be derived.

Serne (ONW 80:4) has determined sorption isotherms for Cs on tuff by column, batch adsorption, and batch desorption methods. From these results a Freundlich isotherm exponent (n) of 0.59 (column) 0.63 (adsorption) or 0.69 (desorption) may be calculated. Batch adsorption isotherms have also been determined for argillite (n = 0.96), basalt (n = 0.82), and granite (n = 0.52).

From data measured by Barney et al (BAR 78) Freundlich isotherm exponents may be calculated for some rocks (concentration range 10^{-5} - 10^{-8} M): Basalt n = 0.85-0.90, granite n = 0.68-0.75, argillite n = 0.85-0.91.

McKinley (MCK 81) has determined sorption isotherms for some geologic materials and obtained the following exponents in a Freundlich isotherm:
Altnabreac weathered granite, $25^{\circ}C$, n = 0.64
Drigg sand (quartz), $25^{\circ}C$, n = 2.01
Camborne Fracture Infill, $25^{\circ}C$ n = 1.2

Rançon (RAN 78) has determined sorption isotherms for river sediments and found that for Cs the isotherm is linear. Saturation of the solid phase is however reached at lower concentration for Cs than for Sr.

Saas (SAA 79) reports a Freundlich isotherm for canal sediments in Rhône water with an exponent of 1.49.

Bo (BO 79) has found a marked dependence on the carrier concentration when sorbing Cs on kaolinite. This dependence levels out at higher concentrations and the author concludes that not even trace amounts of Cs follow linear sorption laws.

Skagius et al (SKA 82:2) have determined sorption isotherms for Cs on granites and fitted the result into Freundlich isotherms with exponents 0.66 (Stripa granite) and 0.54 (Finnsjö granite).

5.2.3 Ionic strength

Erdal et al (ERD 79:5) have measured sorption of Cs on granite and argillite at two different ionic strengths. A decrease with increasing ionic strength was observed for both solids. For two solids, mainly consisting of quartz and clay obtained close to a salt dome, high sorption was observed both in synthetic groundwater and in synthetic brine. Thus a mineral with high capacity and selectivity for Cs seems to be present.

Dosh et al (DOS 78:2) have obtained lower distribution coefficients in brine than in groundwater for dolomite and sandstone. Dosh et al (DOS 79) have also studied the dependence on sodium concentration of Cs-sorption on Culebra dolomite and obtained an equation: K_d = const. $[Na]^{-0.52}$ i.e. an increase in K_d with decreasing sodium concentration.

Lynch et al (LYN 79:2) have observed a faster breakthrough in column studies for Magenta dolomite at higher ionic strength. Lynch et al (LYN 80) also have compared batch and column measured distribution coefficients for Magenta dolomite. In deionized water the difference is quite large - 0.12 m^3/kg in column and 0.65 m^3/kg in batch, while in groundwater approximately the same value was obtained by both methods - 0.019 m^3/kg.

Meyer et al (MEY 79:3) have measured Cs-sorption on Na-montmorillonite at two different ionic strengths: in 0.5 M Na^+ K_d = 0.020 m^3/kg, while in 4 M Na^+ it is 0.005-0.010 m^3/kg.

For SiO_2 the difference is larger (MEY 80:4) - in 5 M NaCl K_d was 0.003 m^3/kg while in 0.01 M it was 0.100 m^3/kg.

Borg et al (BOR 76) reports in a literature survey a distribution coefficient of 0.280 m^3/kg for a basalt in prepared water, while in seawater the value is only 0.007 m^3/kg.

Silva et al (SIL) have determined distribution coefficients on montmorillonite at ionic strengths between $2 \cdot 10^{-3}$ and 1 M. K_d varies from 5 m^3/kg at 0.002 M to 0.010 m^3/kg at 1 M NaCl.

Burkholder et al (BUR 79) report distribuion coefficients for montmorillonite in brine of different concentrations.

Brine concentration (M)	4	2	1	0.5
Distribution coefficient (m^3/kg)	$2 \cdot 10^{-3}$	$5 \cdot 10^{-3}$	$10 \cdot 10^{-3}$	$20 \cdot 10^{-3}$

Relyea et al (REL 77) have determined distribution coefficients for a number of minerals in 5.13 N, and 0.03 N NaCl. The difference in K_d between those is one to two orders of magnitude.

Eichholz et al (EIC 79) have studied Cs-sorption on rocks in different water phases and obtained the following distribution coefficients:

Rock	Distilled water	Equilibrated water	1 N NaCl
Basalt	0.44	0.31	0.06
Shale	1.94	1.15	0.55
Limestone	0.040	0.014	0.017
Dolomite, Magenta	0.91	0.39	0.13
Dolomite, Culebra	1.16	-	0.03

all in m^3/kg.

Thus, there seems to be a pronounced ionic strength dependence.

McHenry (MCH 54) has studied Cs-sorption on Hanford soil and found that only large amounts of salts like sodium nitrate appreciably reduce Cs-sorption.

Saas (SAA 79) has shown that the fixation of Cs on river sediments is 60-70% in river water, while it is only 20-40% in sea water.

Kohtala et al (KOH 79) have shown that for clays the distribution coefficient decreases with increasing ionic strength - from 1 m^3/kg at 10^{-3} M NaCl to 10^{-2} m^3/kg at 1 M NaCl.

5.2.4 Competing ions, and complexation

Meyer et al (MEY 78) observed an increasing K_d with decreasing Ca^{2+}-loading on Na-montmorillonite.

Burkholder et al (BUR 79) report an effort to correlate K_d to a number of factors such as ion exchange capacity, surface area and composition of water phase. The following equation was obtained:
$\ln K_d = 6.9 + 1.7\left[CEC\right] - 0.86\left[SA\right] - 1.7\left[Na\right] + 0.63\left[HCO_3^-\right] - 0.90pH + 0.13Eh + 0.32t$
(K_d is given in ml/g)

McFadden (MCF 80) has compiled data on the influence of organics on nuclear waste nuclides. For Cs very poor complexation with humic acids is observed ($Ce>Fe>Mn>Co\geq Ru\geq Sr>Cs>I$). Competition with other metals in groundwater also reduces Cs-chelate complexation. Cs that has been sorbed on soil has been found not to be removed by hydroacetic acid, EDTA, DTPA, EDDMA or CDTA. Detergents impair Cs-sorption.

Komarneni (KOM 79) has compared sorption on clays in 0.02 N NaCl and 0.02 N $CaCl_2$ and found a higher sorption from NaCl than from $CaCl_2$ solution.

Barney et al (BAR 78:2) have studied sorption of Cs on Hanford sediments and found that the significant variables were: K^+, Na^+, humic acid, and sediment source. Increased concentration of humic acid promotes Cs-sorption. By linear regression the following equation was obtained:
$\ln K_d = 16.8 - 1928\left[K^+\right] - 173\left[Na^+\right] + 0.194\left[\text{humic acid}\right] + 71000\left[K^+\right]^2 + 1000\left[Na^+\right]^2 - 0.005\left[\text{humic acid}\right]^2 + 13000\left[K^+\right]\left[Na^+\right]$
(K_d in ml/g, humic acid in mg/l).

McHenry (MCH 54) has shown that the order of importance of concentration of competing ions on Cs-sorption onto Hanford soil is $H^+>La^{3+}\geq Sr^{2+}>Na^+>NH_4^+$.

Gilham (GIL 80) has found no correlation between major cation concentrations and K_d for Cs on fifteen Canadian clays.

Baudin (BAU 81) has studied the influence of 100 mg/l humic acid on Cs-sorption on illite and attapulgite and found no difference in K_d-values.

Saas (SAA 79) has shown that the order of influence of competing cations on Cs-sorption onto river sediments is $Fe^{2+} > K^{+} > Na^{+} > Ca^{2+} > Mg^{2+}$.

Laske (LAS 79) has compared sorption of Cs on a number of clays in distilled water and $Ca(OH)_2$ or $CaSO_4$ solutions. He has in general obtained very high distribution coefficients and small differences between the water phases.

5.2.5 Ion exchange capacity

Erdal et al (ERD 79:3) have measured distribution coefficients for several samples of tuff from Yucca Mountain and found variations by three orders of magnitude. The high-sorbing samples have a high content of zeolites with high CEC, while the low-sorbing samples contains mainly feldspars and SiO_2, but there are also non-zeolitized samples with high K_d.

Tamura (TAM 60) has found that illite, that has a lower CEC than montmorillonite, is selective for Cs, i.e. has a higher sorption than can be attributed for by CEC. Kaolinite also has a higher Cs-sorption than expected from CEC.

Burkholder et al (BUR 79) report an investigation of the influence of several parameters on Cs-sorption. By linear regression the following equation was obtained:

$$\ln K_d = 6.9 + 1.7\,[CEC] - 0.86\,[SA] - 1.7\,[Na] + 0.63\,[HCO_3^{-}] - 0.90\,pH + 0.13\,Eh + 0.32\,t$$

(K_d in ml/g)

Gilham (GIL 80) has found no discernable correlation between K_d and CEC for fifteen Canadian clay minerals.

Miettinen et al (MIE 81) have not found any correlation between K_d and CEC for some Finnish clays, tills, sediments, and fracture fillings.

5.2.6 Temperature

Erdal et al (ERD 77), (ERD 78), (ERD 79:2) have measured distribution coefficients at 22^{0}C and 70^{0}C for a number of rocks. For some rocks, e.g. for granite, an increse by a factor of 3 was observed while for others, as argillite, devitrified tuff and alluvium, only small changes and in some cases even decreases, were observed. The time to obtain sorption equilibrium for alluvium decreased from 12 days to 3 days when increasing the temperature from 22^{0}C to 70^{0}C (ERD 77).

Kenna (KEN 80) has studied Cs-sorption on subseabed smectite sediments (illite, clinoptilolite, montmorillonite) and found a decrease in distribution coefficient when changing from 20^{0}C to 60^{0}C (from 1.2 to 0.34 m^3/kg).

Means et al (ONW 80:2) have found a very slight increase in K_d for Na-montmorillonite when changing the temperature from 25 to 250^{0}C.

For Cs-sorption on montmorillonite, Burkholder et al (BUR 79) report that there is no influence of a temperature change from 20 to 70^{0}C.

Komarneni (KOM 79) has found a significant decrease in sorption with increasing temperature for montmorillonites, illites, micas, kaolinites, chlorites and shales. For montmorillonite in 0.02 N $CaCl_2$ a decrease in sorption from 1.05 meq/100 g at 25^{0}C to 0.69 meq/100 g at 80^{0}C was found. The change with temperature may be due to a change in hydration radius of Cs.

Komarneni et al (KOM 80) have also studied sorption by clay minerals that have been hydrothermally heated (400^{0}C, 300 bars). Micaeous vermiculites and well crystallized illites, containing vermiculite, were greatly affected by the treatment (changes in K_d by up to two orders of magnitude), while poorly crystallized illites were very slightly affected.

The selectivity of clinoptilolite with respect to Cs-sorption is temperature dependent in the presence of Na^+ (AME 60:2).

Rançon (RAN 78) has measured Cs-sorption on kaolinite, illite and bentonite at $0^{o}C$, $10^{o}C$ and $23^{o}C$ and obtained a decrease in K_d between $0^{o}C$ and $23^{o}C$ by a factor of 1.5.

5.2.7 Contact time

Erdal et al (ERD 79), (ERD 79:2), (ERD 80:2), (WOL 77), (WOL 79) have studied Cs-sorption vs time for a number of solids. The time for obtaining "equilibrium" varies among the solids: for alluvium most of the sorption occurs during the first 10 days, while for bentonite the sorption has almost no time dependence. For argillite equilibrium is reached after 10-30 days. An increase with time is also observed for tuff. For granite there is an increase with time that may be due to changes in mineralogy during the experiment.

Dosh et al (DOS 79:2) have found an increasing K_d with time for Eleana argillite.

Kenna (KEN 80) concludes that sorption equilibrium is established in 3-4 days at $20^{o}C$ (faster at $60^{o}C$) for subseabed smectite sediments.

Beall et al (BEA 78) have observed two time-dependent sorption processes for attapulgite, one fast, that is complete in about 30 min and one slower, that is in equilibrium within one day.

Burkholder et al (BUR 79) report an equation obtained by linear regression:

$$\ln K_d = 6.9 + 1.7[CEC] - 0.86[SA] - 1.7[Na] + 0.63[HCO_3^-] - 0.90pH + 0.13Eh + 0.32t$$

(K_d is given in ml/g)

A small time dependence at contact times between 3 and 47 days has been observed for geologic media.

Barney et al (BAR 78) have studied the sorption kinetics for Cs on granite, basalt, and argillite. A large part of the sorption occurs during the first five minutes, but equilibrium was not reached within 35 days. For still longer contact times (BAR 79) the sorption continued, but a strange effect was observed for basalt, where a sudden desorption occurred at 100-150 days, followed by a continued sorption. The desorption was attributed to formation of colloidal particles on which Cs is sorbed.

Sumrall et al (SUM 68) observes that sorption equilibrium is reached for Yazoo and Zilpa clays in far less than 90 minutes.

McKinley (MCK 81:2) has found that sorption equilibrium for Altnabreac granite was reached in 10 days, while for Drigg sand (quartz) it only took three hours.

5.2.8 Redox Potential (Eh)

Erdal et al (ERD 79:5), (ERD 80:2) have performed sorption measurements in controlled atmosphere (N_2-atm, <0.2ppm O_2, <20ppm CO_2). Small or no differences were observed compared to oxic conditions for granite argillite, tuff and solids obtained from the vicinity of a salt dome ("Aquifer" and "Aquitard", mostly quartz and clay).

The linear regression equation reported by Burkholder et al (BUR 79) shows a quite small Eh-dependence (c.f. 5.2.7); $\ln(K_d)$ is proportional to 0.13 Eh (K_d in ml/g).

Relyea et al (REL 78), (REL 79:5) have compared sorption under oxic and anoxic conditions and found that although Cs is not affected, there are changes in the groundwater/mineral equilibrium that may cause changes in sorption. The largest observed difference in distribution coefficient between oxic and anoxic conditions was one order of magnitude. In most cases no difference was observed.

5.2.9 Liquid/solid, particle size, surface area

Wolfsberg (WOL 77) has found no difference in distribution coefficient between 1 and 2 g of bentonite in the same amount of liquid phase. Erdal et al (ERD 79) have obtained lower distribution coefficients for chunk samples of argillite than for ground samples. For granite, Erdal et al (ERD 80:2) have found a dependence on particle size. Small particles have higher distribution coefficients than larger particles. Wolfsberg et al (WOL 81:3) have found no dependence on solid/liquid ratio for Cs-sorption on tuff. Daniels et al (DAN 81:4) have found higher K_d for particles <38 µm. For particles larger than 38 µm no dependence on particle size was observed.

Dosh et al (DOS 79:2) have studied Cs-sorption on Eleana argillite and seen no particle size dependence in the range 34 µm - 4000 µm.

Burkholder et al (BUR 79) observes that Cs-sorption seems to increase with increasing surface area.

Vandergraaf et al (VAN 82) concludes that for sorption on granitic drill core material the effect of surface roughness was not as great as that of the chemical and mineral composition of the rock.

5.3 Distribution Coefficients

Comparisons of distribution coefficients measured by batch and column methods have been performed in Los Alamos (ERD 79:5), (ERD 80:2), (WOL 81:3), (DAN 82). The following ratios between distribution coefficients measured by the two methods were found.

Tuff	$K_{d,batch}/K_{d,column}$	= 2.4-3.6
Granite	" "	< 0.5
Argillite	" "	< 3

In columns where clay, mica or zeolites are present, a continuous leaking of Cs as well as the peak was observed.

An alternative "batch" method, i.e. circulating columns, has been tried (WOL 81:3) for tuff, giving $K_{d,batch}/K_{d,circ.column} = 0.4\text{-}2.0$.

Lynch et al (LYN 80) have compared batch and column measurements for Magenta dolomite. In groundwater the distribution coefficients are approximately the same for both methods, while for deionized water $K_{d,batch}/K_{d,column} = 5.4$.

Seitz et al (SEI 78) have determined K_d for oolitic limestone by both batch and column methods and obtained $K_d = 0.080\ m^3/kg$ in batch and $K_d = 0.120\ m^3/kg$ in column.

Burkholder et al (BUR 79) compare distribution coefficients for montmorillonite at different ionic strengths, obtained by batch and column methods. There seems to be no major difference between the values obtained in these ways.

Komarneni (KOM 79:2) has found that K_d for mixtures of clay minerals with vermiculite is generally higher than would be expected from K_d-data for the pure minerals.

Brookins (BRO 76), (BRO 78) reports from Oklo that most Cs has migrated.

Champlin (CHA 62) has found in a study of injection of waste solutions into sandstone, that Cs showed a lower retention than Sr, Co and Ru.

Calmels et al (CAL 79) have performed field migration experiments in fissured granite. Their preliminary results imply that Cs is fixed in an almost irreversible manner. The same investigators (GOB 81), however, report a lower K_d for Cs than for Sr in in-situ measurements.

Behrens et al (BEH 81:2) have compared distribution coefficients obtained by batch and column methods for fluviatile and aeolian sands and not found any significant difference.

Neretnieks et al (NER 82) have studied migration of Cs in a single fissure in a drill core and observed a retention that would be expected from batch measured distribution coefficients.

Dlouhy (DLO 81) has compared batch and column determined distribution coefficients for soil: batch K_d = 0.110 m^3/kg, column K_d = 0.065 m^3/kg.

Wolfsberg et al (WOL 82) have studied Cs migration in a column of crushed granite and found that more than half of the Cs was left in the column when the main peak had been eluded. Of this retained Cs 66% was still at the load point of the column.

5.3.1 Rocks

At Los Alamos, sorption studies for a tuffacous alluvium (Frenchman Flats, Nevada Test Site) have been performed (WOL 77), (WOL 78:2), (ERD 77), (ERD 78). Distribution coefficients between 3.5 and 10 m^3/kg have been reported for Cs in synthetic groundwater. For Eleana argillite from Nevada Test Site (major minerals: quartz, sericite, chlorite, albite, siderite and pyrophyllite) distribution coefficents for Cs in the range 0.9-3.7 m^3/kg (4.3 m^3/kg for desorption) have been reported (ERD 78), (ERD 78:2), (ERD 79), (ERD 79:7), (DAN 81:2).

The sorption of Cs in groundwater on Climax Stock granite from the Nevada Test Site has been studied at Los Alamos (ERD 78:2), (ERD 79:2), (ERD 79:7), (ERD 80:3), (DAN 81:2). Distribution coefficients between 0.22 and 0.5 m^3/kg under atmospheric conditions and 1.0-2.7 m^3/kg under anoxic conditions have been obtained. Tuff samples from several sites have been studied in Los Alamos (ERD 78), (ERD 79:3), (ERD 79:7), (WOL 79), (VIN 80), (WOL 80), (ERD 81:2), (DAN 82). The distribution coefficients vary over a wide range - from 0.04 to 42.6 m^3/kg and in some cases variations of three orders of magnitude are found for samples from the same site. In general no high K_d-values are found for devitrified tuffs while high values are found for zeolitized tuffs, but there are exceptions.All measurements in Los Alamos cited above have been performed in groundwater solution.

Dosh et al (DOS 78:2) have measured Cs-sorption on Magenta and Culebra dolomites. In groundwater K_d ranges from 0.004 to 0.010 m^3/kg. For sandstone (Bell Canyon and Cherry Canyon) K_d in groundwater was higher - 0.030 to 0.140 m^3/kg. Dosh et al (DOS 79:2) have also studied sorption on Eleana argillite. In groundwater the distribution coefficient for particles <74 μm was 1.1 to 4.1 m^3/kg. For synthetic dolomites, Dosh et al (DOS 79) have obtained distribution coefficients between 0.0005 and 0.215 m^3/kg in synthetic groundwater. Lynch et al (LYN 79) have studied sorption on argillite and hornfels in synthetic groundwater and obtained K_d 1.3-4.0 m^3/kg for Syncline Ridge argillite (quartz, illite, etc.), 5.1 m^3/kg for Calio Hills argillite and 2.4 m^3/kg for Calio Hills hornfels. Lynch et al (LYN 80) reports a distribution coefficient of 0.019 m^3/kg for Magenta dolomite.

Meyer et al (MEY 78) have measured distribution coefficients for limestone and basalt within the "controlled sample" program. For crushed limestone in groundwater (10^{-3} M Na^+) an average K_d of 0.485 m^3/kg was obtained at pH 8-8.4 and for basalt the value was 0.677 m^3/kg at pH 7.7-8.2.

Borg et al (BOR 76) have compiled K_d-data from several US investigations. As reports on these are not available in this survey some data from Borg et al are given below:

Rock	Water	K_d m^3/kg
Alluvium, Nevada	Groundwater	0.121-3.165
Tuff, Rainier Mesa	Prepared water	1.020
Tuff, Rainier Mesa	Prepared water	12.1-17.8
Tuff, Nevada Test Site	Groundwater	5-8
Carbonate, Yucca Flats	Prepared water	0.014
Granodiorite, Climax Stock	Prepared water	1.0-1.8
Granite, Nevada	Deep formation water	0.035
Basalt, Buckboard Mesa	Prepared water	0.792-9.520
Pillow basalt, Amchitka	Prepared water	0.039
Basalt, Amchitka	Prepared water	0.280
Basalt, Amchitka	Sea water	0.007
Shaley siltstone	Well water	0.309
Sandstone, NM	Well water	0.102

MacLean et al (MAC 78) at LLL report the following distribution coefficients:

Rock	Water	K_d m^3/kg
Limestone	Limestone water	0.060
Limestone	Brine	0.001
Basalt	Basalt water	0.29
Basalt	Brine	0.002
Dolomite	Amargosa water	0.033
Granite, Climax Stock	Granite water	0.380
Granite, Westerly	Granite water	0.210
Tuff	Basalt water	0.122
Conasauga shale	Limestone water	4.43
Green argillaceous shale	Limestone water	8.00

Ramspott et al (RAM 77) report for Climax Stock granite in granite water K_d between 0.46 and 1.5 m^3/kg. Ramspott et al (RAM 79:3) report for Bonanza King Dolomite a K_d of 0.033 m^3/kg.

Seitz et al (SEI 78) have obtained $K_d = 0.080$ m^3/kg for oolitic limestone.

Burkholder et al (BUR 79) report a K_d of 0.30 m^3/kg for Umtanum basalt at pH 8.6 in 0.003 N $NaHCO_3$.

Serne (ONW 80:4) has investigated sorption on 15 different sandstones in brine and obtained distribution coefficients of 0-0.008 m^3/kg. Relyea et al (REL 77) have determined distribution coefficients for some rocks in different ionic media (K_d in m^3/kg):

Rock	5.13 N NaCl	0.03 N $CaCl_2$	0.03 N NaCl
Granite	0.0005	2.67	3.89
Limestone	0.002	1.55	0.712
Basalt	0.006	0.58	0.379

Serne et al (SER 77) have measured K_d for some rocks in distilled water: Magenta dolomite 2.65 m^3/kg, Culebra dolomite 1.74 m^3/kg, "Cherry Canyon No 1" (limestone) 7.51 m^3/kg, "Cherry Canyon No 2" (calcareous sand) 18.56 m^3/kg, "Cherry Canyon No 3" (siliceous sand) 6.75 m^3/kg.

Eichholz et al (EIC 79) have determined distribution coefficients for some rocks in water equilibrated with the rocks: basalt 0.31 m^3/kg, shale 1.15 m^3/kg, limestone 0.014 m^3/kg, Magenta dolomite 0.39 m^3/kg.

Barney et al (BAR 77) reports for some rocks: basalt 0.296 m^3/kg in 0.03 N $NaHCO_3$, granite (Climax Stock) 0.650 m^3/kg in 0.03 N $NaHCO_3$ + 0.06 N $CaSO_4$, and argillite (Eleana) 0.986 m^3/kg in totally 0.03 N Na^+, Mg^{2+}, Na^+ solution.

Komarneni (KOM 79) reports 0.07 m^3/kg for Conasauga Shale in contact with neutralized (pH 7) waste solution with 22.4 mg Cs/l.

Parker et al (PAR 79) report distribution coefficients for some rocks: sandstone 0.102 m^3/kg, siltstone 0.309 m^3/kg, shaly siltstone 0.309 m^3/kg, sediments 0.300 m^3/kg, soils 0.250-1.053 m^3/kg, tuff 18.0 m^3/kg, basalt 0.792-9.52 m^3/kg, granodiorite 0.008-1.81 m^3/kg and granite 0.034 m^3/kg.

Tschurlovits (TSC 79) reports distribution coefficients for some austrian granites: Grein Aumühle, fresh broken 0.40 m^3/kg, Herrschenberg Waldviertel, cutted slab 0.130 m^3/kg, Neustadel Platte, fresh broken 0.90 m^3/kg, Gebharts Waldviertel 1.25 m^3/kg, Neuhaus >2.50 m^3/kg, Kastenhof, weathered surface 0.060 m^3/kg, Grein Aumühle >2.50 m^3/kg, Gloxwald 0.60 m^3/kg.

Carlsson et al (CAR 78) have measured a K_d for granodiorite of 0.25 m^3/kg.

Berak (BER 63) has determined K_d-values for a large number of silicate rocks: rhyolites K_d = 0.004-100.0 m^3/kg, quartz porphyrites K_d = 0.032-0.730 m^3/kg, diabases K_d = 0.014-0.800 m^3/kg and basalts 0.122-0.900 m^3/kg.

5.3.2 Pure minerals

Meyer et al (MEY 80) have determined distribution coefficients for γ-alumina. At pH 6 K_d = 0.003 m^3/kg and at pH 9 K_d = 0.010 m^3/kg. Meyer et al (MEY 80:4) have also measured K_d for trace Cs (10^{-2} M) on SiO_2. In 0.01 M NaCl K_d = 0.100 m^3/kg.

MacLean et al (MAC 78) report for biotite in granite water a distribution coefficient of 0.680 m^3/kg.

Burkholder et al (BUR 79) report for montmorillonite in 0.5 M brine 0.020 m^3/kg, for albite in 0.03 N $NaHCO_3$ 0.53 m^3/kg and for anorthite in 0.3 N $NaHCO_3$ 0.610 m^3/kg.

Relyea et al (REL 77) have measured distribution coefficients for a number of minerals in different ionic media (equilibrium pH, 10 days contact time, values in m^3/kg):

Mineral	5.13 N NaCl	0.03 N $CaCl_2$	0.03 N NaCl
Illite	0.502	53.03	43.01
Montmorillonite	0.032	0.341	1.22
Vermiculite	0.466	82.41	7.18
Biotite	1.168	87.91	64.7
Kaolinite	0.005	2.49	0.750
Quartz	0.0009	0.042	0.057
Albite	0.017	0.861	0.397
Anorthite	0.015	0.612	0.450
Microcline	0.007	0.223	0.342
Hornblende	0.008	3.10	0.872
Enstatite	0.001	0.218	0.115
Augite	0.003	0.509	0.228

Komarneni (KOM 79:2) has determined distribution coefficients for some natural zeolites at pH 7 in a diluted and neutralized waste solution with 22.4 mg Cs/l:

Clinoptilolite	, AZ	1.07 m^3/kg
-"-	, ID	1.14 -"-
Mordernite	, AZ	2.88 -"-
-"-	, NV	2.02 -"-
Erionite	, OR	2.02 -"-
-"-	, CA	2.53 -"-
Chabazite	, AZ	1.83 -"-
Phillipsite	, CA	2.02 -"-

In this study some clays etc. were included:

Montmorillonite, AZ	0.12 m^3/kg
Vermiculite , SC	0.46 -"-
Ferric oxide	0
Manganese(IV)dioxide	0
Gibbsite	0

Komarneni et al (KOM 80) have also studied some vermiculite containing clays.

Cooper micaceous vermiculite	1.43 m^3/kg
Wisconsin illite	0.007 -"-
Goose Lake illite	0.041 -"-
Fithian illite	0.100 -"-
Green illite	0.031 -"-
Beavers Bend illite	0.206 -"-
Indian Red Clay	0.225 -"-

Baudin (BAU 81) report for illite at pH 7 a distribution coefficient of 1.3 m^3/kg , and for attapulgite at the same pH 0.16 m^3/kg.

Rançon (RAN 79) has measured K_d for a number of clays in synthetic granitic groundwater at equilibrium pH:

Attapulgite (pH 7.5)	4 m^3/kg
Bauxite (pH 7)	0.15 m^3/kg
Bentonite (pH 10)	1.5 m^3/kg
Clinoptilolite (pH 6.5)	9.0 m^3/kg
Illite (pH 7.5)	4.0 m^3/kg
Kaolinite (pH 6.5)	0.2 m^3/kg
Sepiolite (pH 8.5)	10.0 m^3/kg
Vermiculite (pH 8.5)	11.0 m^3/kg

Rançon (RAN 67) reports for clays in natural water (pH 8.5): illitic clays 21.0 m^3/kg, kaolinitic clays 0.30 m^3/kg, and montmorillonites 2.0 m^3/kg.

Skytte-Jensen (SKY 80) report K_d values for some Danish clays: kaolinite 0.32 - 3.2 m^3/kg, illite 10 - 32 m^3/kg, and montmorillonite 1 - 100 m^3/kg, depending on ionic strength (10^{-3} - 10^{-1} M, Na^+ and Ca^{2+} varied).

Pushkarev et al (PUS 62) have studied sorption on manganese oxide (MnO_2) in distilled water and not observed any sorption over a wide pH-range.

Carlsson et al (CAR 78) report for chlorite K_d=0.02 m^3/kg.

Bonhole et al (BON) report sorption data for vermiculite that corresponds to distribution coefficients of 0.88 - 5.4 m^3/kg, depending on particle size.

Doshi et al (DOS 73) have studied sorption/coprecipitation of Cs with manganese dioxide and obtained very low or no sorption.

Berak (BER 63) has performed a thorough study of Cs-sorption on a large number of silicate minerals and rocks and report the following results:

Silicate group	Example of minerals	K_d m^3/kg
Tectocilicates	quartz	0 - 0.061
-"-	zeolites	0.009 - 25.0
Nesosilicates	olivine	0 - 0.054
Sorosilicates	epidote	0.006
Cyclosilicates	beryl	0 - 0.008
Inosilicates	pyroxenes, amphiboles	0 - 0.900
Phyllosilicates	micas, kaoline, chlorite	0 - 4.9

Furnica et al (FUR 73) have studied sorption on minerals in distilled water and obtained the following distribution coefficients:

Mineral	K_d m^3/kg
Montmorillonite	8
Diatomite	3
Kiselguhr	1.1
Silicagel	0.092
Kaoline	0.42
Marble	0.005

Lee (LEE 73) reports for natural vermiculite 0.027 - 0.082 m^3/kg, for K^+ saturated vermiculite 0.152 - 0.325 m^3/kg and for Na^+ saturated vermiculite 0.220 - 0.988 m^3/kg.

5.3.3 Soils, clays and sediments

For bentonite distribution coefficients between 1.6 and 2.0 m^3/kg have been measured in groundwater by Wolfsberg et al (WOL 77).

Brat et al (BRA 73) report K_d values of 13 - 50 m^3/kg for vermiculite at pH 9, while Chandra (CHA) has obtained only 0.29 - 0.64 m^3/kg at the same pH. This autor, however, report a higher K_d for bentonites - 9 - 14 m^3/kg at pH 9.

Erickson (ERI 79) has found a K_d of 0.100 to 10.0 m^3/kg for depending on concentration.

Kenna (KEN 80) has measured a K_d of 1.2 m^3/kg for subseabed smectite sediments.

Nowak (NOW 80) has found very low distribution coefficients (0.0004 - 0.005 m^3/kg) for bentonite and hectorite in brine.

Beall et al (BEA 78) have obtained a distribution coefficient of 1.0 m^3/kg for attapulgite in 0.25 M NaCl solution.

Brockett et al (BRO 53) have determined the distribution coefficient for Cs on Conasauga shale in tap water at different solid/liquid ratios. In all cases K_d was >10 m^3/kg.

Meyer et al (MEY 78) have determined K_d for attapulgite in 0.25 M Na^+ at pH 5 to 0.876 m^3/kg, and at 4 M Na^+ to 0.089 m^3/kg. For Na-montmorillonite K_d varied from 7.6 to 16.3 m^3/kg, depending on Ca^{2+}-loading. Meyer et al (MEY 79:3) have obtained a K_d of 0.020 m^3/kg for Na-montmorillonite in 0.5 M Na^+.

Soratesn et al (SOR 60) have obtained a K_d of about 200 m^3/kg for illite.

Rançon (RAN 78) report for sediments in Rhône water K_d >10 m^3/kg.

Kohtala et al (KOH 79) have measured distribution coefficients for some clays in synthetic groundwater and obtained for a heavy clay 0.723 m^3/kg and for a muddy clay 0.188 m^3/kg.

Miettinen et al (MIE 81) have determined K_d values for five different heavy clays (K_d = 4.4, 4.7, 0.6, 1.4, and 18.0 m^3/kg respectively). one silty clay (K_d = 6.0 m^3/kg) and four different sandy tills (K_d = 0.56, 2.1, 0.20, and 0.56 m^3/kg respectively) in groundwater and at 10^{-6} - 10^{-12} M Cs.

Baetsle et al (BAE 62) have determined K_d in tap water at pH 7.7 for some sands: sand with organic content 0.038 m^3/kg, yellow aeolian sand 0.089 m^3/kg, white sand of Mol 0.030 m^3/kg.

Laske (LAS 79) has measured sorption on a number of clays and marls in distilled water as well as in $Ca(OH)_2$ and $CaSO_4$ solutions:

Solid	K_d m^3/kg		
	dist. water	$Ca(OH)_2$	$CaSO_4$
"Opalinus"clay[1]	10	30	10
"Molasse" clay A[2]	20	20	20
-"- B[3]	>20	30	50
-"- C[4]	50	30	30
"Bänder" clay[5]	20	20	10
"Keuper" clay[6]	50	90	80
"Bolus" clay[7]	3	1	2
Kaoline	0.51	0.27	0.20
Marl "Felsenau"[8]	40	30	30

Main minerals in order of occurrence:
1 = illite, kaolinite, chlorite
2 = montmorillonite, illite, kaolinite
3 = montmorillonite, quartz, illite
4 = illite, calcite, chlorite, quartz
5 = calcite, illite, quartz, feldspar
6 = corrensite, illite, quartz
7 = kaolinite, Fe-oxide/hydroxide, quartz
8 = dolomite, illite/muscovite, vermiculite, quartz

Dlouhy (DLO 67) has measured distribution coeffecients for two Italian soils - Cassaccia tuff (K_d = 0.8 - 1.0 m^3/kg) and Cassaccia soil (K_d = 0.16 - 0.20 m^3/kg)

Galea et al (GAL 65:2) reports for Na-bentonite in distilled water K_d = 0.027 m^3/kg at pH 7.5, 0.050 m^3/kg at pH 9 and 0.030 m^3/kg at pH 11.

5.4 Suggested sorption mechanisms.

Jacobs (JAC 62) has observed that Cs-sorption by illite-type clay minerals (non-expanding 2:1 layer lattice) is seriously inhibited by cations, having steric favourableness for edge fixation sites. K^+, Ru^+, and NH_4^+ usually inhibit Cs-sorption. In freely expanding clay minerals simple ion-exchange occurs. For vermiculite interlayer fixation occurs when the Cs-concentration is increased. This is due to a reduction of the (001) lattice spacing to about 10 Å, causing a physical entrapment of exchangeable cations.

Meyer et al (MEY 79:3) have found that for Cs-sorption on Na-montmorillonite at low loading the reaction cannot be explained only by ion exchange as the slope of K_d vs $[Na^+]$ is not -1.

The sorption isotherms for Cs on bentonite may be fairly well predicted by the MINEQL-program (SIL 79:4).

Komarneni et al (KOM 79:3) have investigated the degree of fixation of Cs by shales and shale minerals and found that 91 % of the sorbed Cs was fixed by vermiculite and not desorbed by 0.1 N KCl.

Brown (BRO 67) has found that in column leaching of Hanford sediments only 11 % of the sorbed Cs was removed by 50 column volumes of water and the following 500 column volumes desorbed an additional 4 %. The Cs that was leached could easily be resorbed on soil.

Komarneni et al (KOM 80) have shown that the Cs-fixation by clay minerals increased upon hydrothermal heating.

Barney et al (BAR 78) have tested the reversibility of Cs-sorption and observed that a fast desorption occurs for basalt and argillite, while the desorption from granite is much slower.

Ames (AME 60:2) has found that the zeolite clinoptilolite is highly Cs-selective over wide pH, flow rate, and temperature ranges. The selectivities affected by cation size, charge, electronic structure, and in the presence of Na, temperature.

According to Sawhney (SAW 72) clays are selectively sorbing cations with low hydration energy due to interlayer collapse. These ions (Cs^+, K^+, Rb^+, NH_4^+, etc) are therefore fixed in interlayer positions. The interlayer collapse is also dependent on the layer charge. Desorption may be done by adding ions with high hydration energy (Ca^{2+},Mg^{2+}) which are causing interlayer expansion.

Hawkins et al (HAW 65) also concludes that illitic clay is selective for Cs.

Vandergraaf et al (VAN 82) observes that when sorbing Cs on drill core material from the Canadian shield, an enhanced sorption is obtained on mafic minerals and on chlorite infilling.

Carlsen et al (CAR 81:2) have studied sorption on kaolininte and observed an unexpectedly low K_d at very low salt concentrations. This may be explained by degradation of the clay structures and possibly also by competition from colloidal material.

Levi et al (LEV 67) have studied diffusion of Cs in vermiculite and observed that no movement of Cs could be detected in Cs saturated clay, while trace amounts of Cs are mobile.

Berak (BER 63) concludes that silicate minerals possesses an enhanced sorption affinity for Cs compared to Sr. Only secondary minerals (weathering products) are active sorbents. For many silicate minerals a high K_d for trace amounts is accompanied by a low sorption capacity. Well developed structure changes (bentonitization) leads to high capacity minerals. However, in one ryolithe studied both high sorption capacity and high K_d for trace Cs was observed.

Sawhney (SAW 64) observes that micas and vermiculite sorb more Cs in the presence of Ca^{2+} and similar divalent cations than in the presence of K^+ and similar monovalent cations. Montmorillonite, on the other hand sorbed more Cs in the presence of K^+ than in the presence of Ca^{2+}. While micas and vermiculite fixed a large proportion of the Cs sorbed, montmorillonite and kaolinite did not fix significant amounts of Cs. It is suggested that the fixation of Cs by micas occurs at the crystal edges, where Cs may replace K and thus become a part of the stable mica structure.

Lee (LEE 74) has studied the sorption of Cs by vermiculite treated with Na-ion. Here the role of Na is to induce a wider c-axis spacing in which Cs ion can be sorbed. CEC and K_d seems to be influenced by the c-axis spacing.

6. SORPTION OF TECHNETIUM

6.1 Chemistry of technetium

In aqueous solution Tc forms TcO_4^- under oxidizing conditions but upon reduction reverts to sparingly soluble Tc(IV) (BON 81).

At neutral pH there is a strong tendency to hydrolysis under reducing conditions. Few complex formers react with Tc, but polyalcohols (present close to plant roots) may cause complexation (MCF 80).

In soils containing appreciable amounts of organic matter, Tc(VII) may be reduced to Tc(IV) and be adsorbed (JOH 80:2).

EH-pH calculations for the geochemistry found in Oklo predict retention of Tc at 25°C and migration at 200°C (BRO 78).

A review of Tc-chemistry is given in a Prav report (TOR 81).

6.2 Parameters of importance for sorption

According to Barney (BAR 78:2) none of the parameters identified to be importance for sorption of Sr and Cs are important for Tc-sorption.

6.2.1 pH

Meyer et al (MEY 80) have observed a pH-dependence for TcO_4^--sorption on ZrO_2 from NaCl-solutions at low ionic strenghts. At 0.01 M NaCl the sorption was decreased by a factor of 20 between pH 2 and 10. Meyer et al (MEY 80:2) have also observed similar effects for Al_2O_3.

Burkholder et al (BUR 79) report a linear regression equation giving K_d for Tc proportional to -1.3pH (K_d in ml/g) when not taking time into account and K_d proportional to -0.41pH with time dependence.

6.2.2 Ionic strength

Dosh (DOS 78) reports the same low K_d (<0.001 m^3/kg) for both brine and groundwater.

Meyer et al (MEY 80) report an increase in K_d for ZrO_2 when decreasing the ionic strength, especially at low pH, and a similar effect for Al_2O_3 (MEY 80:2).

Relyea et al (REL 77) observes positive K_d values in brine for minerals that show no sorption in groundwater.

The sorption also seems to be salt dependent , as Relyea et al (REL 79:4) have observed an increase in K_d with decreasing salt concentration only after 30 d contact time.

6.2.3 Competing ions, and complexation

Burkholder et al (BUR 79) report a linear regression equation where K_d is proportional to -0.86 $[Ca]$, +1.2 $[Cl]$, -0.61 $[HCO_3]$ without time dependence and to -0.20 $[Ca]$, +0.67 $[Cl]$, -0.50 $[HCO_3]$ with time dependence (K_d in ml/g).

Bird (BIR 79) has observed a TcO_4^- sorption by lead minerals.

6.2.4 Ion exchange capacity

Burkholder et al (BUR 79) report a linear regression equation where K_d for Tc is proportional to -0.8 [CEC] without time dependence and to -0.52 [CEC] with time dependence (CEC = cation exchange capacity)

Relyea et al (REL 79:4) have not obtained any correlation between CEC and K_d for Tc.

6.2.5 Temperature

Erdal et al (ERD 79) have observed a drastical decrease in TcO_4^--sorption when the temperature is raised from 22°C to 70°C , but the same autors (ERD 79:2) have also observed increases in K_d with temperature and concludes that there is no consistent temperature effect. Increases with a factor of 5 - 10 in K_d for quartz monzonite and argillite have been obtained (ERD 79), (ERD 79:8).

Francis et al (FRA 79:3) have shown that increasing temperature enhances the initial reduction rate of TcO_4^- but the rate exponentially decayed at high temperatures.

6.2.6 Contact time

Erdal et al (ERD 80:2) have observed a strong time dependence for Tc-sorption on argillite under controlled atmosphere conditions. It may be due to a slow release of a reducing agent. Daniels et al (DAN 81:2) reports that ^{95m}Tc in reduced form that was added to areated groundwater was still in reduced form after one week. Erdal et al (ERD 81:2) report high K_d values (0.03 - 0.1 m^3/kg) for bournonite and tetrahedrite after a contact ime of one week, but almost zero after six weeks. Crowe et al (CRO 80:2) have studied sorption on Yucca Mountain tuff and observed an increasing sorption rate with time. Also for Eleana argillite an increase with time (up to 59 days was studied) was observed by Erdal et al (ERD 79).

For basalt, granite and shale, Francis et al (FRA 79:3) have observed an increase in TcO_4^--loss with time.

Burkholder et al (BUR 79) report a K_d proportional to -0.09 [t], using a linear regression equation.

Relyea et al (REL 77) report an increase in K_d for granite, basalt, and limestone after 30 d contact time.

6.2.7 Redox potential (Eh)

Erdal et al (ERD 80:2) report higher K_d under controlled atmosphere conditions (i e low O_2 content) for granite, tuff, and argillite. Wolfsberg et al (WOL 80) have obtained higher K_d values for tuff in N_2 atmosphere than in air. Crowe et al (CRO 80:2) report 10 - 100 times higher K_d values for tuff under controlled atmosphere (<2 ppm O_2, <20 ppm CO_2) than in air.

Francis et al (FRA 79:3) have shown that addition of Fe^{2+} prior to TcO_4^- addition did not affect the reduction, but addition of Fe^{2+} to Tc-rock suspensions gives a reduction. The reduced form of Tc is slowly oxidized to TcO_4^- under oxic conditions. Losses of TcO_4^- from basalt/water mixtures occurs at Eh 0.240 V higher than theoretically necessary for TcO_2 formation.

Burkholder et al (BUR 79) report a linear regression expression where K_d for Tc is proportional to -1.57 Eh without time dependence and to -0.93 Eh with time dependence.

Relyea et al (REL 78) also observes that Tc-adsorption is increased by removal of oxygen from the system, although K_d was not always correlated with the measured Eh.

Dosh et al (DOS 80) concludes that Tc-sorption on argillite from Calico Hills was associated with the presence of elemental carbon capable of reducing Tc.

Strickert et al (STR 80) observes a sorption of TcO_4^- by powdered lead, while sulfides and oxides with the metal ion in the higher ocidation states are not effective.

6.2.8 Solid/liquid, particle size, surface area

Crowe et al (CRO 80:2) have found little particle size dependence for Tc-sorption on Yucca Mountain tuff in controlled atmosphere. Erdal et al (ERD 79) however observes a correlation between surface area and K_d for ground samples of Eleana argillite, while the highest K_d was obtained for chunk samples. For Climax Stock granite, Erdal et al (ERD 79:2) have found a sorption on a finely ground fraction (K_d= 0.031 - 0.080 m^3/kg) while there was no sorption for larger fractions.

Burkholder et al (BUR 79) report a linear regression expression where K_d is proportional to -0.31 SA without time dependence and to -0.22 SA with time dependence (K_d in ml/g, SA = surface area m^2/kg).

Relyea et al (REL 79:4) have observed that K_d for Tc does not correlate with the surface area as K_d for Sr does.

6.3 Distribution coefficients

In Oklo a migration of Tc is observed and hot aqueous fluids have transported Tc about 10 m from the reactor (BRY 76), (GAN 79).

Around a waste deposit in Hanford, Tc has been traced in groundwater up to 15 miles from the disposal site (BRO 67).

Strickert et al (STR 80) compares batch and column measurements for some rocks and minerals:

Solid	K_d m^3/kg Column	Batch
LASL tuff	<0.001	<0.001
Limestone	0.002	<0.001
Buornonite	0.40-0.77	0.20-2.0
Chalcopyrite	<0.001	<0.001
Galena	3	0.001-0.010

6.3.1 Rocks

Daniels et al (DAN 81:2) report for granite (Climax Stock) K_d = 0.004 - 0.120 m^3/kg and for Eleana argillite K_d = 0.17 - 0.83 m^3/kg. For tuff Wolfsberg et al (WOL 80) report lower values: 0.0002 - 0.0003 m^3/kg for devitrified and zeolitized tuff in air and 0.008 - 0.026 m^3/kg in N_2-atmosphere. Erdal et al (ERD 79) report for Eleana argillite 0.011 - 0.031 m^3/kg. All values in synthetic groundwater.

Dosh et al (DOS 79) report for dolomites in synthetic groundwater: synthetic dolomite K_d = 0, Culebra dolomite K_d = 0.001, and Magenta dolomite K_d = 0.0015 m^3/kg. Lynch et al (LYN 79) have obtained the following K_d-values in deionized water: argillite (Syncline Ridge) 0.016 - 0.196 m^3/kg, argillite (Calico Hills) 0.155 - 6.5 m^3/kg and hornfels (Calico Hills) 0, all under oxic conditions.

Ramspott et al (RAM 79) observes that Tc is not sorbed on oolitic limestone or basalt in groundwater and that K_d = 0 for Climax Stock granite in groundwater (RAM 79:3).

Coles et al (COL 79) have not obtained any sorption on St Peter sandstone in dynamic experiments, and Weed et al (WEE 81) have obtained similar results for a sandstone core.

Strickert et al (STR 80) have obtained K_d <0.001 m^3/kg for apatite, CR basalt, chalk, NTS dolomite, PNL granite, granite, LT metamorphic, Pierre shale, PNL shale, mottled siltstone, LASL tuff, NTS tuff, and PNL limestone. For PNL basalt K_d = 0.001 m^3/kg.

Relyea et al (REL 77) have observed a $K_d \neq 0$ at long contact times (30 d) for argillaceous shale, Westerly granite, Sentenniel Gap basalt, and oolitic limestone. Serne et al (SER 77) have obtained low K_d-values for Culebra dolomite (0.0002 - 0.0005 m^3/kg) and Cherry Canyon limestone (0.002 - 0.007 m^3/kg) in groundwater and brine.

Eichholz et al (EIC 80) report for shale (pH 4.8 - 4.9 and equilibrated water) 0.0008 - 0.001 m^3/kg, basalt 0.0008 - 0.001 m^3/kg, limestone 0.0009 - 0.001 m^3/kg and for sand 0.0011 - 0.0099 m^3/kg.

6.3.2 Pure minerals, oxides

Erdal et al (ERD 81:2) report for bournonite 0.10 m^3/kg, boulangerite 0.088 m^3/kg and tetrahedrite 0.028 m^3/kg.

Nowak (NOW 80) report for hectorite 0.001 m^3/kg in 5 N NaCl brine.

Meyer et al (MEY 80) report for TcO_4^- sorption on ZrO_2 on 0.01 M NaCl-solution K_d = 0.600 m^3/kg at pH 2 and 0.03 m^3/kg at pH 10. For Al_2O_3 the same autors (MEY 80:2) report K_d = 0.005 m^3/kg at pH 9.

Tewhery et al (TEW 78) have not obtained any sorption on biotite in groundwater.

Strickert et al (STR 80) report the following K_d values for some minerals etc:

Mineral	K_d m^3/kg
Bornite	<0.003
Bournonite	0.2-2.0
Chalcopyrite	<0.001
Chrysocolla	0.01-0.02
Enargite	<0.001
Galena	0.001-0.010
Pyrite	<0.003
Tennantite	<0.003
Tetrahedrite	0.002-0.016
CdS	0.0003
Cu_2S	0.00099
CuS	0.00004
FeS	0.00099
HgS	0.00002
PbS	0.00046
ZnS	0.00008

Relyea et al (REL 77) have obtained $K_d = 0$ in groundwater for illite, kaolinite, montmorillonite, vermiculite, biotite, quartz, albite, anortite, microcline, hornblende, enstatite, and augite. Relyea et al (REL 78) also report the following pH, Eh and K_d values for some minerals in 5.13 N brine and 10 days contact time:

Mineral	pH	Eh mV	K_d m^3/kg
Illite	7.2	570	1.49
Montmorillonite	8.3	677	0
Vermiculite	8.2	632	0.35
Biotite	8.3	709	0
Kaolinite	6.8	654	0.013
Quartz	6.8	654	0.013
Albite	7.8	386	6.12
Anortite	8.6	718	2.7
Microcline	7.0	693	0.05
Hornblende	8.1	695	0.59
Enstatite	8.1	718	0.27
Augite	8.8	720	0.81

In 0.03 N NaCl the following values were obtained:

Mineral	pH	Eh mV	K_d m^3/kg
Illite	7.4	250	17.0
Montmorillonite	8.9	568	0
Vermiculite	8.0	613	0.002
Biotite	8.3	480	0
Kaolinite	6.2	464	0
Quartz	6.5	251	0.002
Albite	8.4	551	5.38
Anorthite	8.6	643	1.09
Microcline	7.6	542	0.008
Hornblende	8.7	539	0.008
Enstatite	8.6	656	0.010
Augite	8.7	416	1.6

Rançon (RAN 79) report distribution coefficients for TcO_4^- in synthetic groundwater: bauxite 0, laterite 0, limonite 0.006 m^3/kg, goethite 0.009 m^3/kg, siderite 0.060 m^3/kg, chalcopyrite 0, cerusite 0.002 m^3/kg, and galena 0.040 - 3.0 m^3/kg.

6.3.3 Soils, clays, sediments

Erickson (ERI 79:2) report for abyssal red clay in 0.68 N NaCl and oxidizing conditions $K_d = 0$.

Dosh et al (DOS 78:2) have obtained some sorption of Tc on clay from a halite core.

Francis et al (FRA 77) report from soil chhromatography a K_d for Muscatine Silt Loam of 0.00009 m^3/kg at 0.01 meq Ca/l and for Fuquay sand 0.00002 m^3/kg.

Routson et al (ROU 77) observes that K_d for a South Carolina subsoil is essentially 0.

Nowak (NOW 80) has not obtained any Tc-sorption on bentonite in > 5 N NaCl brine.

6.4 Suggested sorption mechanisms

For Tc the sorption obviously occurs due to a reduction of TcO_4^-, followed by a precipitation of $TcO_2(s)$, which has been observed by many investigators (c. f. for instance FRA 79:3). The reduction may be due to the presence of reducing agents as Fe^{2+}, Fe(s) or Pb(s), sulfide or oxide minerals with the metals in the lower oxidation states (STR 80).

7. SORPTION OF IODINE

7.1 Chemistry of Iodine

In aqueous solutions iodide is the predominating species over a wide pH range, but in highly basic solutions iodate (IO_3^-) is formed (JOH 80:2).

In natural waters, especially surface waters and in soils the iodide is to a large extent converted by microbial action (BEH 81). The iodide is oxidized and incorporated in organic compounds, probably proteins. Also iodate is affected and reduced. This effect is more pronounced in soil than in surface water, In surface fresh water the formed organic radioiodide compounds are mainly in solution while in soils they are to a larger extent bound to insoluble organic substances.

7.2 Parameters of importance for sorption

7.2.1 pH

Rançon (RAN 79:2) has observed that K_d for iodide on minerals decreases with increasing pH.

Kepak (KEP 79) has obtained a maximum in iodide sorption on hydrated ferrous, ferric, aluminum, and chromic oxides at pH 4 - 7.

7.2.2 Nuclide concentration

McLellon (MCL 67) has studied the concentration dependence for iodine sorption on hydrous ferric oxide and obtained a good fit to a Freundlich isotherm at 10^{-13} - 10^{-7} M.

DeRegge et al (DER 79) have measured the sorption of iodide on Boom clay and on glauconite sand and obtained the same (low) distribution coefficients from 0.1 ppm I^- to 1000 ppm, i e no concentration dependence has been observed in this concentration range.

7.2.3 Competing ions, complexation, and chemical reaction

Strickert et al (STR 80) have observed a high sorption for iodide and iodate on copper, iron, and lead sulphide minerals and concludes that iodine is sorbed by minerals whose metal ions form insoluble iodides or iodates. Iodine is also sorbed by silver powder and this suggests that a reaction incorporating the iodine into the lattice occurs.

Bird et al (BIR 79) have found that iodide sorption is obtained for Cu-metal at a Eh suitable for CuO_2 formation. They have, however, observed that lead minerals like massicot and litharge (PbO) sorb I, while cerussite ($PbCO_3$) has no sorption.

Rançon (RAN 79) observes that the best I-sorption is obtained for minerals containing cations capable of forming sparingly soluble iodides.

7.2.4 Contact time

Strickert et al (STR 80) have studied the sorption kinetics for Cu_2S and bournonite. A fast initial sorption was obtained, but still after 100 h equilibrium was not reached.

7.2.5 Solid/liquid, particle size, surface area

Strickert et al (STR 80) have changed the amount of sorbent from 0.1 to 0.4 g and not observed any change in K_d. Neither the particle size seems to be important.

7.3 Distribution coefficients

In Oklo the iodine has migrated (BRY 76) and in situ measurements by Calmels et al (CAL 79) show no measurable retention of iodide.

7.3.1 Rocks

Dosh (DOS 78) has obtained K_d <0.001 m^3/kg for dolomite in both brine and groundwater.

Johnston et al (JOH 80) report $K_d = 0$ for both devitrified and zeolitized tuff.

Borg et al (BOR 76) have obtained $K_d = 0.001$ m^3/kg for Rainier Mesa tuff.

Fried et al (FRI 78) have performed a large investigation of iodide sorption on rocks and pure minerals in preequilibrated water and report the following values for rocks:

Rock	K_d m^3/kg
Basalt	0.002
Chalk	0.005
Dolomite	0
Granite	0.001
Siltstone	0.005
Shale	0.002
Los Alamos tuff	0.008
NTS tuff	0.001

Daniels et al (DAN 81:2) have measured iodine sorption from a solution prepared by leaching a rock sample from an underground nuclear test with a synthetic water. K_d values at pH 8.6 - 8.7: Eleana argillite 0.002 - 0.006 m^3/kg, alluvium 0.020 m^3/kg, tuff 0.

Strickert et al (STR 80) have performed a large investigation of sorption of iodide and iodate and report the following K_d values:

Rock K_d m^3/kg:	I^-	IO_3^-
Apatite	<0.001	
CR basalt	0.002	<0.001
PNL basalt		<0.001
Chalk	<0.001	0.005
NTS dolomite	<0.001	<0.001
Granite	<0.001	<0.001
PNL granite		<0.001
Pierre shale	<0.001	0.001
PNL shale		0.009
Mottled siltstone	<0.001	0.005
LASL tuff	<0.001	<0.001
NTS tuff	<0.001	<0.001
PNL limestone		0.016

Meyer et al (MEY 79) have obtained K_d = 0 for Culebra dolomite, Magenta dolomite and green argillaceous shale in 0.01 M NaCl at pH 5.6 - 7.6.

7.3.2 Pure minerals, oxides

Meyer et al (MEY 79) have measured iodide sorption in 0.01 M NaCl at pH 5.6 - 7.6. K_d = 0 was obtained for montmorillonite, anhydrite, oolitic limestone, augite and hornblende.

Fried et al (FRI 78) report the following distribution coefficients for iodide in preequilibrated water:

Mineral	K_d m^3/kg
Apatite	0.0008
Anhydrite	0.005
Galena	0.052
Tetrahedrite	>0.330
Bornite (with molybdenite)	0.163
Bournonite	>0.330
Chalcopyrite	>0.330
Pyrite	>0.330
Chrysocolla	0.063
Enargite	>0.330
Tennantite	0.108
Azurite	0.052 - 0.108
Bornite	0.108 - 0.163
Enargite	>0.330
Al_2O_3	0
Cu_2S	>0.330
CuS	0.330
CdS	0.108
HgS	0.008
PbS	0.108
ZnS	0
CuO	0
PbO_2	0
Fe_2O_3	0.005
CuCl	0.004
$FeCl_2$	0.001
$FeSO_4$	0
Ag (powder)	>0.330

Erdal et al (ERD 81:2) report K_d values in tuff groundwater (all in m^3/kg): bournonite: 0.140, boulangerite: 0.140, tetrahedrite: 0.

Strickert et al (STR 80) have measured sorption of iodide and iodate on a number of minerals:

Mineral	K_d m^3/kg: I^-	IO_3^-
Anhydrite		0.004
Witherite		0.001
Bornite (Cu_5FeS_4)	0.030-0.970	0.010-0.100
Bournonite ($CuPbSbS_3$)	0.200-1.000	0.960
Chalcopyrite ($CuFeS_2$)	0.190	0.400-1.300
Chrysocolla ($CuSiO_3$ $2H_2O$)	0.003-0.012	0.040-0.075
Enargite (Cu_3AsS_4)	0.130-0.150	0.070-0.120
Galena (PbS)	0.026	0.003-0.045
Pyrite (FeS_2)	0.30-3.00	>1.00
Tennantite ($((Cu,Fe)_{12}As_4S_{13})$)	0.100-0.35	0.010-0.020
Tetrahedrite ($((Cu,Fe)_{12}Sb_4S_{13})$)	1.10-1.50	1.0-3.0
-"- ($((Cu,Ag,Fe)_{12}Sb_4S_{13})$)	0.004-0.68	0.032

Bird et al (BIR 79) have studied iodide sorption on minerals and found that augite, hornblende and vermiculite had no affinity for I^-, while biotite had some, although spontaneous desorption occurred quickly. Lead minerals like massicot and litharge (PbO) sorbs I^-, while cerussite ($PbCO_3$) has so sorption. Most of these measurements were performed at pH 7 - 8.

Rançon (RAN 79) concludes that I^- is poorly or not at all sorbed on aluminum silicate rocks and report the following K_d values:

Mineral	K_d m^3/kg	pH
Bauxite	0.001	7.5
Laterite	0.0002	7.5
Limonite	0.006	6.7
Goethite	0.001	7.7
Siderite	0.0002	7.3
Chalcopyrite	0.215	7.8
Crysocolla	0.010	7.2
Cerrusite	0.003-0.008	7.7
Galena	0.015-0.050	7.0-7.5

Relyea et al (REL 77) report K_d = 0 in 0.03 N $CaCl_2$, NaCl and $NaHCO_3$ at pH 7 for illite, kaolinite, montmorillonite, vermiculite, biotite, quartz, albite, anorthite, microcline, hornblende, enstatite, and augite.

Rochon et al (ROC 79) report for quartz K_d = 0, illite K_d = 0.001 and vermiculite K_d = 0.001 m^3/kg.

Kepak (KEP 79) has observed a maximum K_d of about 0.01 m^3/kg for hydrated oxides.

Furnica et al (FUR 73) report sorption data for some solids: montmorillonite 0.001 m^3/kg, diatomite 0.002 m^3/kg, kiselguhr 0.003 m^3/kg, silica gel 0.001 m^3/kg, and kaoline 0.001 m^3/kg, all in distilled water.

7.3.3 Soils, clays, sediments

Meyer et al (MEY 79) have obtained K_d = 0 for Conasauga shale.

Dosh et al (DOS 78:2) report distribution coefficients in the range 0 - 0.004 m^3/kg for clay from a halite core in brine or groundwater.

Brockett et al (BRO 53) have obtained distribution coefficients between 0.026 and 0.071 m^3/kg for Conasauga shale in tap water.

In soil chromatograms, Francis et al (FRA 77) have obtained a K_d = 0 for Muscatine Silt Loam and K_d 0.0002 m^3/kg for Fuquay sand.

DeRegge et al (DER 79) have obtained 0.001 - 0.005 m^3/kg for Boom clay and glauconite sand.

7.4 Suggested sorption mechanisms

Anionic iodide and iodate may be sorbed by minerals with high anion exchange capacity, but the best sorption is obtained by minerals, containing metal ions, capable of forming insoluble iodides or iodates (STR 80). For iodide ions, coprecipitaiton or incorporation into the crystal lattice may be incolved and sorption of iodate seems to coincide with iodide sorption (STR 80).

Cations in solution, capable of forming sparingly soluble iodides also gives a sorption (precipitation) (RAN 67).

8. CONCLUSIONS

Although the distribution coefficients for a mineral/rock differ due to properties of solids as well as due to different methodologies at the laboratories, some general conclusions may be drawn on the sorptive behaviour for the studied fission products.

8.1 Cobalt

The hydrolysis of cobalt seems to be the prime factor determining the sorption. This means that pH is of great importance and a maximum in sorption is observed at a certain pH, where easily sorbed hydrolysis products are formed. Decreases in sorption at higher pH are usually ascribed to formation of neutral or negatively charged hydrolysis products, which are poorly sorbed at the negative mineral surfaces.

Parameters like nuclide concentration, CEC, and temperature are of minor importance. For clays the sorption usually is fast (minutes), while for rocks much longer times (weeks) may be required for complete sorption.

8.2 Nickel

Chemically, nickel is very similar to cobalt. An evaluation of the pH-dependence for sorption has not been possible with the avalible data. A hydrolysis related sorption may be expected for Ni as well, although the pH where sorption becomes significant is higher than for cobalt.

CEC and temperature are of minor importance also for nickel.

8.3 Strontium

Trace strontium is not hydrolysed in groundwater at naturally occurring pH, but at pH above 9 carbonate complexation may become important, and cause an increase in sorption.

A low but significant sorption with more or less pronounced pH-dependence is observed for most minerals, especially for clays, where distribution coefficients of 1 - 5 m^3/kg may be encountered. The pH-dependence seems to be due to the nature of the solid, rather than to Sr-chemistry.

The concentration and temperature dependencies are in most cases neglible, while the ionic strength dependence is strong, as well as the CEC-dependence.

As the only valence state of Sr in water solution is +II, Eh is not important for Sr properties. The only influence on sorption of Eh is thus due to changes in the mineral.

A non-selective sorption due to electrostatic interactions between the negatively charged mineral surface and the Sr^{2+} ions seems to occur. This is interpreted a "ion exchange" by many investigators although there is no evidence of the presence of sorption sites of different selectivity.

There are also some evidence that Sr is enriched in calcite rich soils, although calcite is a poor Sr-sorbent. This is probably due to dissolution/precipitation processes.

8.4 Cesium

Cesium is not hydrolyzed or complexed in groundwater.

The pH-dependence for cesium sorption is usually minor, although a sligth increase with increasing pH may be encountered, especially for clays.

A strong concentration dependence is often encountered and Freundlich sorption isotherms for cesium may have exponents between 0.2 and 2.0. Usually the exponent is <1.

A selectivity for Cs-sorption has been observed for some minerals, usually layer silicates and zeolites.

Ionic strength may be important, but a correaltion between K_d and CEC is not always evident.

Temperature, and particle size are of minor importance.

The sorption of cesium on rocks and minerals may be divided into one fast sorption and a slower, usually interpreted as a diffucion into the rock matrix. The main part of the sorption is due to the fast mechanism.

As the only valence state for Cs in water solution is +I, the only effect of Eh on sorption may be changes in the solid phase.

8.5 Technetium

Technetium is anionic under oxidizing conditions in groundwater and a sligth sorption on minerals with high AEC may be observed, especially at low pH.

Under anoxic conditions (low Eh) Tc is precipitated and thus sorbed.

The time dependence of the sorption seems to be related to the kinetics of the Eh-controlling reactions.

8.6 Iodine

Iodine is anionic in groundwater and slightly sorbed by minerals having a high AEC.

A high sorption of iodine is observed for minerals, containing anions capable of forming iodides with low solubility (Hg, Pb, Ag, etc).

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Table I. Distribution coefficients for Cobalt. Low and medium ionic strength, pH 7-9.

Rock	Origin/type	Water	K_d m^3/kg	Reference
Granite	Altnabraec	synth. g.w.	0.024-0.491	MCK 81:2
Fracture fillings	Finland	-"-	2.9-9.7	MIE 81
Alluvium		nat. g.w.	0.013	WOL 78:2
Sand		synth. g.w.	0.001	KOH 79
Sand	Drigg	-"-	0.001	MCK 81:2
Sand			0.040	DLO 81
Clay				
Montmorillonite			0.1-3.1	SKY 80
Kaolinite			0.003-0.3	-"-
Illite			1-10	-"-
Soil, etc.				
Conasauga shale		tap w.	70	MEA 78
Different clays	Scandinavia	synth. g.w.	0.042-2.24	KOH 79
Soil	Finland	-"-	0.264	-"-
Heavy clay		synth. g.w.	0.41-14	MIE 81
Silty clay		-"-	4.5	-"-
Sandy till		-"-	0.14-0.88	-"-

TABLE II. Distribution coefficients for nickel. Low and medium ionic strength, pH 7-9.

Solid	Water	K_d m^3/kg	Reference
Fuquay sand	0.01 meq/ml Ca	0.005	FRA 77
Muscatine silt loam	-"-	>0.015	-"-
Conasauga shale		0.0065	FRA 78
Na-montmorillonite	pH 6.3	0.0036	MEY 79:6

TABLE III. Distribution coefficients for strontium. Low and medium ionic strength, pH 7-9.

Rock	Origin/type	Water	K_d m^3/kg	Reference
Granite	Climax Stock	g.w.	0.004-0.030	LASL[a]
-"-	- " -	g.w.	0.059	MAC 78
-"-	- " -	g.w.	0.005	BAR 77
-"-	quartz monzonite	g.w.	0.020	ERD 78
-"-	Granodiorite CS	prep. w.	0.004-0.023	BOR 76[b]
-"-	Nevada	deep form. w.	0.002	BOR 76[b]
-"-	Westerly	g.w.	0.007	MAC 78
-"-		0.03 N $NaHCO_3$	0.03	REL 77
-"-			0.006	BAR 79
-"-		g.w.	0.003	MAY 79
-"-		dist. w.	0.003-0.015	TSC 79
-"-	quartz porphyrite	river w.	0.006-0.018	BER 63
Syenite		g.w.	0.015	MAY 79
Gabbro		g.w.	0.008	MAY 79
Anorthosite		g.w.	0.025	MAY 79
Sandstone	New Mexico	well w.	0.001	BOR 76[b]
-"-	sediments, TX		0.140-0.180	ONW 80:2
-"-		dist. w.	0.019-0.038	SER 77
Diabase		river w.	0-0.045	BER 63
Rhyolite		river w.	0-0.056	BER 63
Basalt	Nevada Test site	prep. w.	0.016-0.135	BOR 76[b]
-"-	Pillow, Amchitka	- " -	0.220	BOR 76[b]
-"-	Amchitka	- " -	1.220	BOR 76[b]
-"-		g.w.	0.108	MEY 78
-"-		-"-	0.045	MAC 78
-"-		0.03 N $NaHCO_3$	0.150	BUR 79
-"-		- " -	0.1	REL 77
-"-			0.041-0.092	SER 79
-"-		g.w.	0.151	BAR 77
-"-		- " -	0.100	BAR 79
-"-		river w.	0.006-0.187	BER 63
Argillite	Eleana	g.w.	0.029-0.200	LASL[a]
-"-	- " -	g.w.	0.021-0.060	DOS 79:2
-"-	- " -	- " -	0.191	BAR 77
-"-		- " -	0.130	ERD 78
-"-		- " -	0.040-0.070	LYN 79
-"-		- " -	0.108	BAR 79

Table III 2(5)

Rock	Origin/type	Water	K_d m^3/kg	Reference
Tuff	Yucca Mountain	g.w.	0.041-25	LASL[a]
-"-	Jackass Flats	- " -	0.035-45	LASL[a]
-"-	G-Tunnel, NTS	- " -	39-54	LASL[a]
-"-	vitric	- " -	13	ERD 78
-"-	devitrified	- " -	0.060	ERD 78
-"-	zeolitized	- " -	0.240	ERD 78
-"-	Ranier Mesa	prep. w.	0.260-4.3	BOR 76[b]
-"-	Nevada Test Site	g.w.	0.400	BOR 76[b]
-"-	- " -		4.0	BOR 76[b]
-"-		g.w.	0.040	MAC 78
-"-	Casaccia		0.045-0.075	DLO 67
Alluvium		g.w.	0.200	ERD 78
-"-		g.w.	0.048-2.45	BOR 76[b]
Silt	Muscatine Silt Loam	0.1 meq/ml Ca^{2+}	0.001-0.004	FRA 77, 78, 78:2
-"-	Shaley Siltstone, NM	well w.	0.008	BOR 76[b]
Shale	Conasauga	tap w.	0.63	BRO 53
-"-	- " -	g.w.	0.029	MAC 78
-"-	Green argillaceous	g.w.	0.060	MAC 78
-"-		0.03 N $NaHCO_3$	0.4	REL 77
Hornfels		g.w.	0.020	LYN 79
Limestone	Oolitic	g.w.	0.004-0.007	SEI 79
-"-		- " -	0.005-0.006	MEY 78, 78:3
-"-		- " -	0.003	MAC 78
Limestone		0.03 N $NaHCO_3$	10.0	REL 77
-"-		dist. w.	0.010	SER 77
-"-			0.001-0.015	SER 79
-"-	"Calcareous rock"		0.014	RAN 77
-"-	marble		0.002-0.010	FUR 73
Dolomite	Magenta, Culebra	g.w.	0.001-0.032	SAND[c]
-"-	Rustler Formation	prep. w.	0.005-0.014	BOR 76[b]
-"-		g.w.	0.001	MAC 78
-"-		dist. w.	0.006-0.012	SER 77
-"-		synth. g.w.	0.150-0.180	MIE 81

a = experiments described in a series of reports from LASL: ERD 77, WOL 77, ERD 78, ERD 78:2, ERD 78:3, WOL 78, ERD 79, ERD 79:2, ERD 79:3, ERD 79:4, ERD 79:5, ERD 79:6, ERD 79:8, ERD 79:9, WOL 79, CRO 80:2, ERD 80, ERD 80:2, ERD 80:3, VIN 80, VIN 80:2, WOL 80, DAN 81:2, DAN 81:3, DAN81:3, DAN 81:4, ERD 81:2, WOL 81:3)

b = Values quoted from a literature survey. Original publications not available in this survey.

Table III 3(5)

c = experiments described in a series of reports from Sandia: DOS 78, DOS 78:2, DOS79, DOS79:2, HIN 77

Mineral group	Mineral	Water	K_d m^3/kg	Reference
Nesosilicates			0-0.035	BER 63
Sorosilicates			0.014	BER 63
Cyclosilicates			0-0.015	BER 63
Inosilicates			0-0.030	BER 63
-"-	augite	0.03 N $NaHCO_3$	0.8	REL 77
-"-	enstatite	- " -	0.1	REL 77
-"-	hornblende	- " -	0.3	REL 77
Phyllosilicates			0-0.217	BER 63
-"-	biotite	0.03 N $NaHCO_3$	0.6	REL 77
-"-	kaolinite		0.036-0.055	FUR 73
Tectosilicates			0-0.210	BER 63
-"-	quartz	0.03 N $NaHCO_3$	0.3	REL 77
-"-	- " -	synt. g.w.	0.070-0.090	ROC 79
-"-	albite	0.3 N $NaHCO_3$	0.004	BUR 79
-"-	- " -	0.03 N $NaHCO_3$	0.5	REL 77
-"-	anorthite	0.3 N $CaCl_2$	0.003	BUR 79
-"-	- " -	0.03 N $NaHCO_3$	0.4	REL 77
-"-	microcline	- " -	0.4	REL 77
-"-	clinoptilolite		0.088-0.750	KOM 79
-"-	- " -		0.102	AME 62:3
-"-	mordernite		0.282-0.583	KOM 79
-"-	- " -		0.038	AME 62:3
-"-	erionite		0.019-0.025	KOM 79
-"-	chabazite		0	KOM 79
-"-	- " -		0.028	AME 62:3
-"-	phillipsite		0	KOM 79
-"-	stilbite		0.008	AME 62:3
-"-	analcite		0.006	AME 62:3
-"-	sodalite		0.004	AME 62:3
Oxides	alumina	0.01 M Na^+	1.0	MEY 80
-"-	- " -		> 10	RAN 67
-"-	ferric oxide (red)		0.079	KOM 79
-"-	ferric oxide (yellow)		3.9	KOM 79
-"-	manganese dioxide		2.0	KOM 79
-"-	titanium dioxide		0.1	DOS 77
-"-	gibbsite		0	KOM 79
Carbonates	calcite		0.030	RAN 78

Table III 4(5)

Clay	Water	K_d m^3/kg	Reference
Montmorillonite		1.25-1.46	BRO 53
-"-	0.1 M NaCl	0.080	ORNL[a]
-"-	0.03 N $NaHCO_3$	1.2	REL 77
-"-		0.010-32	SKY 80
Kaolinite		0.43	BRO 53
-"-		4	SOR 60
-"-		0.015	TAM 62
-"-	0.03 N $NaHCO_3$	0.2	REL 77
-"-	0.05 N NaCl	0.011	REL 79:4
-"-		0.1	RAN 79
-"-		0.010-0.310	SKY 80
Halloysite		0.80	BRO 53
Montronite		1.25	BRO 53
Vermiculite		0.95	BRO 53
-"-	0.03 N $NaHCO_3$	1.0	REL 77
-"-	0.05 N NaCl	0.260	REL 79:4
Vermiculite		0.2	RAN 79
-"-		0.010	ROC 79
Attapulgite	0.005 M Na^+	1.9	ORNL[a]
-"-		0.004	BAU 81
-"-		0.70	RAN 79
Illite	0.005 M Na^+	13.01	ORNL[a]
-"-	0.03 N $NaHCO_3$	0.5	REL 77
-"-		0.140	BAU 81
-"-		1	RAN 79
-"-		0.2-0.3	ROC 79
-"-		0.004-1.0	SKY 80
Bauxite		0.1	RAN 79
Clinoptilolite		1	RAN 79
Sepiolite		0.15	RAN 79
Type of Bentonite			
"Commercial drilling mud"	g.w.	0.900-2.90	ERD 77, WOL 77
Metabentonite		0.60-0.75	BRO 53
		4	RAN 79
MX-80	-"-	0.604-1.315	ERI 81:2
Eberslöh	-"-	0.421-0.549	ERI 81:2

a = experiments described in a series of reports from ORNL: MEY77:2, MEY78, MEY78:3, MEY79, RAF81, SMI79

Soil	Water	K_d m^3/kg	Reference
Abyssal red clay	0.68 N Nacl	0.128	ERI 79
Smectite sediments		0.1-0.2	KEN 80
Weathered Conasauga Shale		0.008-0.058	SPA 80
Hanford Soil		0.050	BRO 67
Oak Ridge Clay		4-5	LAC 54
Savannah River soil		0.001-0.010	RYA 81
Clay (diff. Finnish, Danish, Norweigian)	g.w.	0.174-0.245	KOH 79
Clay (diff. Finnish soils)	-"-	0.004-0.110	MIE 81
Sand, Brunkeberg		0.019	CAR 78
Clay, - " -		0.076	CAR 78
Silt, Porsön, Luleå		0.026	CAR 78
- " -		0.004	CAR 78
Sand, Porsön Luleå		0.003	CAR 78
Trombay soil	g.w.	0.143-0.315	GOD 67
Casaccia soil		0.006-0.015	DLO 67
Sandy soil		0.013	DLO 81

TABLE IV. Distribution coefficients for cesium. Low and medium ionic strength, pH 7-9.

Rock	Origin/type	Water	K_d m^3/kg	Reference
Granite	Climax Stock, Nevada	synth. g.w.	0.4-0.5	ERD 79:2
-"-		anoxic w.	1.0-2.6	ERD 80:3
-"-	Climax Stock, Nevada	anoxic w.	0.75-2.7	DAN 81:2
-"-	-"-	synth. g.w.	0.34-0.42	-"-
-"-	-"-	g.w.	0.22-0.51	ERD 78:2
-"-		-"-	0.32-0.55	ERD 79:7
-"-	Nevada	deep form. w.	0.035	BOR 76
-"-	granodiorite	synth. w.	1.0-1.8	-"-
-"-	Climax Stock, Nevada	granite w.	0.380	MAC 78
-"-	Westerly	-"-	0.210	-"-
-"-	Climax Stock, Nevada	-"-	0.46-1.5	RAM 77
-"-		0.03 N $CaCl_2$	2.67	REL 77
-"-		0.03 N $NaCl$	3.89	-"-
-"-	Climax Stock, Nevada	$NaHCO_3+CaSO_4$(0.09 N)	0.650	BAR 77
-"-	Granodiorite		0.008-1.81	PAR 79
-"-			0.304	-"-
-"-	Altnabreac	synth. g.w.	1.0	MCK 81:2
-"-	Grein, Aumühle		0.40	TCH 79
-"-	Herschenberg		0.130	-"-
-"-	Neustadel		0.9	-"-
-"-	Gebharts		1.25	-"-
-"-	Neuhaus		>2.50	-"-
-"-	Kastenhos(weath.)		0.060	-"-
-"-	Grein, Aumühle		>2.50	-"-
-"-	Gloxwald		0.600	-"-
-"-	Granodiorite		0.25	CAR 78
Monzonite	Quartz	synth. w.	0.440	ERD 78
Porphyrite	-"-	synth. river w.	0.032-0.070	BER 63
Rhyolite		-"-	0.004-100.0	-"-
Sandstone	Bell Canyon	g.w.	0.030-0.140	DOS 78:2
-"-	Cherry Canyon	-"-	-"-	-"-
-"-	New Mexico	well w.	0.102	BOR 76
-"-	Calcareous, Cherry C.	dist. w.	18.56	SER 77
-"-	Siliceous, Cherry C.	-"-	6.75	-"-
Basalt	Buckboard Mesa	synth. w.	0.792-9.52	BOR 76
-"-	Pillow, Amchitka	-"-	0.039	-"-
-"-	Amchitka	-"-	0.280	-"-
-"-	-"-	sea w.	0.007	-"-

Table IV 2(8)

Rock	Origin/type	Water	K_d m^3/kg	Reference
Basalt		basalt w.	0.29	MAC 78
-"-		10^{-3} N Na	0.38-0.68	MEY 78
-"-		0.03 N $CaCl_2$	0.58	REL 77
-"-		0.03 N NaCl	0.38	-"-
-"-		dist. w.	0.44	EIC 79
-"-		-"-, deoxygen.	0.47	-"-
-"-		equil. w.	0.31	-"-
-"-	Umtanum Flow	0.03 N $NaHCO_3$	0.296	BAR 77
-"-			0.792-9.52	PAR 79
-"-		synth. river w.	0.122-0.900	BER 63
Diabase		-"-	0.014-0.800	-"-
Argillite		synth. g.w.	0.9-3.5	ERD 79
-"-		anoxic w.	1.0-2.3	DAN 81:2
-"-	Eleana	g.w.	2.5	ERD 78
-"-		-"-	1.6-2.5	ERD 78:2
-"-		-"-	1.9-3.6	ERD 79:7
-"-	Eleana	-"-	1.1-4.1	DOS 79:2
-"-	Syncline Ridge	-"-	1.3-4.0	LYN 79
-"-	Calio Hills	-"-	5.1	-"-
-"-	Eleana	0.03 N (Na,Mg,Ca)	0.986	BAR 77
Tuff	Zeolitized	nat. g.w.	6.0	WOL 79
-"-	Vitric	-"-	0.150	-"-
-"-		-"-	0.227-21.9	ERD 79:3
-"-	Yucca mountain		0.24-42.6	VIN 80
-"-	Devitrified	nat. g.w.	0.150-0.870	WOL 80
-"-	Zeolitized	-"-	8.6-29.0	-"-
-"-	USW, G1		0.04-0.28	ERD 81:2
-"-	G-tunnel, NTS		20.7-28.6	DAN 82
-"-	USW, G1		0.7-15.0	-"-
-"-	Vitric	g.w.	15.0	ERD 78
-"-	Devitrified	-"-	0.120	-"-
-"-	Zeolitized	-"-	0.600	-"-
-"-	Vitric		6.0	ERD 79:7
-"-	Devitrified		0.15	-"-
-"-	Zeolitized		0.74	-"-
-"-	Rainier Mesa	synth. g.w.	1.02	BOR 76
-"-	-"-	-"-	12.1-17.8	-"-
-"-	Nevada Test Site	g.w.	5-8	-"-
-"-		basalt w.	0.122	MAC 78

Rock	Origin/type	Water	K_d m^3/kg	Reference
Tuff			18.0	PAR 79
-"-	Casaccia, Italy		0.8-1.0	DLO 67
Alluvium		synth. g.w.	3.5-10	WOL 77
-"-		g.w.	8.3	ERD 77
-"-	Frenchman Flats		2.5	ERD 78
-"-	Nevada	g.w.	0.121-3.165	BOR 76
Silt	Shaley siltstone, NM	well w.	0.309	BOR 76
-"-	-"-		0.309	PAR 79
Shale	Conasauga shale	neutr. waste soln.	0.07	KOM 79:2
-"-	-"-	limestone w.	4.43	MAC 78
-"-	Green argillaceous	-"-	8.00	-"-
-"-		dist. w.	1.94	EIC 79
-"-		-"-, deoxygen.	0.94	-"-
-"-		equil. w.	1.15	-"-
-"-		0.03 N $CaCl_2$	16.04	REL 77
-"-		0.03 N NaCl	14.99	-"-
Hornfels	Calio Hills	g.w.	2.4	LYN 79
Limestone	Oolitic		0.080	SEI 78
-"-		limestone w.	0.060	MAC 78
-"-		10^{-3} N Ca	0.485	MEY 78
-"-		0.03 N $CaCl_2$	1.55	REL 77
-"-		0.03 N NaCl	0.712	-"-
-"-	Cherry Canyon	dist. w.	7.51	SER 77
-"-		-"-	0.04	EIC 79
-"-		-"-, deoxygen.	0.25	-"-
-"-		equil. w.	0.014	-"-
Marble		dist. w.	0.005	FUR 73
Dolomite	Magenta, Culebra	g.w.	0.004-0.010	DOS 78:2
-"-	Synthetic	-"-	0.025-0.076	DOS 79
-"-	Magenta	-"-	0.0005-0.215	-"-
-"-	Culebra	-"-	0.0012-0.068	-"-
-"-	Magenta	-"-	0.019	LYN 80
-"-		Amargoza w.	0.033	MAC 78
-"-	Bonanza King		0.033	RAM 79:3
-"-	Magenta	dist. w.	2.65	SER 77
-"-	Culebra	-"-	1.74	-"-
-"-	Magenta	-"-	0.91	EIC 79
-"-	-"-	-"- , deoxygen.	0.90	-"-
-"-	-"-	equil. w.	0.39	-"-
-"-	Culebra	dist. w.	1.16	-"-
-"-	-"-	-"- , deoxygen.	0.93	-"-

Table IV 4(8)

Rock	Origin/type	Water	K_d m^3/kg	Reference
"Carbonate"	Yucca Flats	synth. w.	0.014	BOR 76
Fracture filling	Turku	synth. g.w.	0.97-3.6	MIE 81
Quartz sand	Drigg	-"-	0.07	MCK 81:2
Mineral group	Mineral			
Nesosilicates		synth. river w.	0-0-054	BER 63
Sorosilicates		-"-	0.006	-"-
Cyclosilicates		-"-	0-0.008	-"-
Inosilicates		-"-	0-0.900	-"-
-"-	augite	0.03 N $CaCl_2$	0.509	REL 77
-"-	-"-	0.03 N NaCl	0.228	-"-
-"-	enstatite	0.03 N $CaCl_2$	0.218	-"-
-"-	-"-	0.03 N NaCl	0.115	-"-
-"-	-"-	0.03 N $CaCl_2$, anoxic	0.029	REL 78
-"-	-"-	0.03 N NaCl , -"-	0.041	-"-
-"-	-"-	0.03 N $NaHCO_3$, -"-	0.050	-"-
-"-	augite	0.03 N $CaCl_2$, -"-	0.096	-"-
-"-	-"-	0.03 N NaCl , -"-	0.101	-"-
-"-	-"-	0.03 N $NaHCO_3$, -"-	0.075	-"-
-"-	hornblende	0.03 N $CaCl_2$, -"-	0.117	-"-
-"-	-"-	0.03 N NaCl , -"-	0.427	-"-
-"-	-"-	0.03 N $NaHCO_3$, -"-	0.459	-"-
-"-	-"-	0.03 N NaCl	0.872	REL 77
-"-	-"-	0.03 N $CaCl_2$	3.10	-"-
Phyllosilicates		synth. river w.	0-4.9	BER 63
-"-	biotite	0.03 N $CaCl_2$	87.9	REL 77
-"-	-"-	0.03 N NaCl	64.7	-"-
-"-	-"-	0.03 N $CaCl_2$, anoxic	13.0	REL 78
-"-	-"-	0.03 N NaCl , -"-	5.7	-"-
-"-	-"-	0.03 N $NaHCO_3$, -"-	2.6	-"-
-"-	-"-	granite w.	0.680	MAC 78
-"-	chlorite		0.2	CAR 78
Tectosilicates	(quartz etc)	synth. river w.	0-0-061	BER 63
-"-	SiO_2	0.01 M NaCl	0.100	MEY 80:4
-"-	quartz	0.03 N $CaCl_2$	0.042	REL 77
-"-	-"-	0.03 N NaCl	0.057	-"-
-"-	-"-	HCO_3^- g.w.	0.039	REL 77:2
-"-	-"-	0.03 N $CaCl_2$	0.004	REL 78
-"-	-"-	0.03 N NaCl	0.007	-"-
-"-	-"-	0.03 N $NaHCO_3$	0.009	-"-

Table IV 5(8)

Mineral group	Mineral	Water	K_d m^3/kg	Reference
Tectosilicates	albite	0.03 N $NaHCO_3$	0.53	BUR 79
-"-	-"-	0.03 N $CaCl_2$	0.86	REL 77
-"-	-"-	0.03 N NaCl	0.40	-"-
-"-	-"-	HCO_3^- g.w.	0.377	REL 77:2
-"-	-"-	0.03 N $CaCl_2$, anoxic	0.059	REL 78
-"-	-"-	0.03 N NaCl	0.089	-"-
-"-	-"-	0.03 N $NaHCO_3$	0.106	-"-
-"-	anorthite	0.03 N $NaHCO_3$	0.610	BUR 79
-"-	-"-	0.03 N $CaCl_2$	0.612	REL 77
-"-	-"-	0.03 N NaCl	0.450	-"-
-"-	-"-	HCO_3^- g.w.	0.422	REL 77:2
-"-	-"-	0.03 N $CaCl_2$, anoxic	1.89	REL 78
-"-	-"-	0.03 N NaCl , -"-	2.57	-"-
-"-	-"-	0.03 N $NaHCO_3$, -"-	0.659	-"-
-"-	microcline	0.03 N $CaCl_2$	0.223	REL 77
-"-	-"-	0.03 N NaCl	0.342	-"-
-"-	-"-	HCO_3 g.w.	0.445	REL 77:2
-"-	-"-	0.03 N $CaCl_2$, anoxic	0.065	REL 78
-"-	-"-	0.03 N NaCl , anoxic	0.187	-"-
-"-	-"-	0.03 N $NaHCO_3$, -"-	0.082	-"-
-"-	zeolites	synth. river w.	0.009-25.0	BER 63
-"-	clinoptilolite, AZ	neutr. waste soln.	1.07	KOM 79:2
-"-	-"- , ID	-"-	1.14	-"-
-"-	-"-	synth. g.w.	8.0	RAN 79
-"-	mordernite, AZ	neutr. waste soln.	2.88	KOM 79:2
-"-	-"- , NV	-"-	2.02	-"-
-"-	erionite, OR	-"-	2.02	-"-
-"-	-"- , CA	-"-	2.53	-"-
-"-	chabasite, AZ	-"-	1.83	-"-
-"-	phillipsite	-"-	2.02	-"-
-"-	sepiolite	synth. g.w.	6.0	RAN 79
Oxides	alumina		0.010	MEY 80
-"-	ferric oxide	neutr. waste soln.	0	KOM 79:2
-"-	manganese(IV)oxide	-"-	0	-"-
-"-	gibbsite	-"-	0	-"-
-"-	bauxite	art. granitic w.	2.5	RAN 79
Miscellanceous	diatomite	dist w.	3	FUR 73
-"-	kiselguhr	-"-	1.1	-"-
-"-	silicagel	-"-	0.092	-"-

Table IV 6(8)

Clay	Water	K_d m^3/kg	Reference
Montmorillonite (Na)		7.6-16.3	MEY 78
-"-	0.5 M Na	0.020	MEY 79:3
-"-	0.03 N NaCl	1.00	BUR 79
-"-	0.03 N $CaCl_2$	0.341	REL 77
-"-	0.03 N NaCl	1.22	-"-
-"-	HCO_3^- g.w.	1.036	REL 77:2
-"-	0.03 N $CaCl_2$, anoxic	0.296	REL 78
-"-	0.03 N NaCl , -"-	0.848	-"-
-"-	0.03 N $NaHCO_3$, -"-	0.933	-"-
-"- , AZ	neutr. waste soln.	0.12	KOM 79:2
-"-	natural w.	2.0	RAN 67
-"-	10^{-3}-10^{-1} M Ca+Na	1-100	SKY 80
-"-	dist w.	8	FUR 73
Hectorite (Li-montmorillonite)	brine	0.005	NOW 80
Kaolinite	0.03 N $CaCl_2$	2.5	REL 77
-"-	0.03 N NaCl	0.75	-"-
-"-	0.03 N $CaCl_2$, anoxic	0.080	REL 78
-"-	0.03 N NaCl, anoxic	0.099	-"-
-"-	0.03 N $CaCl_2$, -"-	0.109	-"-
-"-	HCO_3^- g.w.	0.475	REL 77:2
-"-	sea w.	0.004	RAN 78
-"-	synth. g.w.	0.3	RAN 79
-"-	dist. w.	0.51	LAS 79
-"-	10^{-1}-10^{-3} M Ca+Na	0.32-3.2	SKY 80
-"-	dist. w.	0.42	FUR 73
Vermiculite	0.03 N $CaCl_2$	4.99	REL 78
-"-	0.03 N NaCl	6.44	-"-
-"-	0.03 N $NaHCO_3$	3.86	-"-
-"-	HCO_3^- g.w.	4.59	REL 77:2
-"-	synth. g.w.	9.0	RAN 79
-"-		0.88-5.4	BON
-"-		13-50	BRA 73
-"-		0.29-0.64	CHA
-"-		0.027-0.082	LEE 73
-"-	0.03 N $CaCl_2$	82.41	REL 77
-"-	0.03 N NaCl	7.18	-"-
-"- , Cooper micaceous	0.001 N $CaCl_2$	1.43	KOM 80
-"- , SC	neutr. waste soln.	0.46	KOM 79:2

Table IV 7(8)

Clay	Water	K_d m^3/kg	Reference
Attapulgite	0.25 M Na	1.0	BEA 78
-"-	-"-	0.876	MEY 78
-"-		0.22	BAU 81
-"-	synth. g.w.	3.5	RAN 79
Illite		200	SOR 60
-"-	0.03 N $CaCl_2$	53.03	REL 77
-"-	0.03 N NaCl	43.01	-"-
-"-	0.03 N $CaCl_2$, anoxic	5.92	REL 78
-"-	0.03 N NaCl, -"-	9.21	-"-
-"-	0.03 N $NaHCO_3$	7.08	-"-
-"- , Wisconsin	0.001 N $CaCl_2$	0.007	KOM 80
-"- , Green	-"-	0.031	-"-
-"- , Goose Lake	-"-	0.041	-"-
-"- , Fithian	-"-	0.100	-"
-"- , Beavers Bend	-"-	0.206	-"-
Illitic clays	nat. w.	21.0	RAN 67
Illite	sea w.	0.350	RAN 78
-"- du Cantal		1.3	BAU 81
-"-	synth. g.w.	7.0	RAN 79
-"-	10^{-3}-10^{-1} M Ca+Na	10-32	SKY 80
-"-	HCO_3^- g.w	4.19	REL 77:2
Bentonite			
"Commercial drilling mud"	synth. g.w.	1.6-2.0	WOL 77
Bentonite	sea w.	0.265	RAN 78
-"-	synth g.w.	0.2	RAN 79
-"-		9-14	CHA
Na-bentonite	dist w.	0.050	GAL 65:2
Soil etc.			
Conasauga shale		>10	BRO 53
Indian red clay	0.001 N $CaCl_2$	0.225	KOM 80
Sediments		0.300	PAR 79
Soils		0.250-1.053	-"-
Kaolinitic clays	natural w.	0.3	RAN 67
"Opalinus" clay	dist. w.	10	LAS 79
"Molasse" clay A	-"-	20	-"-
-"- B	-"-	>20	-"-
-"- C	-"-	50	-"-

Table IV 8(8)

Soil, etc.	Water	K_d m^3/kg	Reference
"Bänder" clay	-"-	20	-"-
"Keuper" clay	-"-	50	-"-
"Bolus" clay	-"-	3	-"-
Marl "Felsenau"	-"-	40	-"-
Heavy clay	synth. g.w.	0.723	KOH 79
Muddy clay	-"-	0.188	-"-
Heavy clay, Loviisa	-"-	4.4-4.7	MIE 81
-"- , Olkiluodo	-"-	0.6-1.4	-"-
-"- , Salo	-"-	18.0	-"-
Silty clay, Jämsä	-"-	6.0	-"-
Sandy till, Loviisa	-"-	0.56	-"-
-"- , Olkiluodo	-"-	2.1	-"-
-"- , Juuka	-"-	0.20	-"-
-"- , Partala	-"-	0.56	-"-
Sand, with organics	tap w.	0.038	BAE 62
-"- , yellow aeolian	-"-	0.089	-"-
-"- , white of Mol	-"-	0.030	-"-
Soil, Casaccia Italy		0.16-0.20	DLO 67
-"-		0.110	DLO 81
Clay, sandy	10^{-4} M comp. ions	0.3	-"-
-"-	10^{-1} M -"-	0.003	-"-

TABLE V. Distribution coefficients for technetium. Low and medium ionic strength, pH 7-9.

Rock	Origin/type	Water	K_d m^3/kg	Reference
Granite	Climax Stock	g.w.	0.004-0.120	DAN 81:2
-"-	-"-	oxic g.w.	0.031-0.080	ERD 79:2
-"-	Quartz monzonite	-"-	<0.080	ERD 78
-"-	Climax Stock	g.w.	0	RAM 79:3
-"-	-"-	granite w.	0	TEW 78
-"-	Westerly	-"-	0	-"-
-"-		preeq. w.	<0.001	STR 80
-"-	PNL granite	-"-	<0.001	-"-
Basalt	CR	-"-	<0.001	-"-
-"-	PNL	-"-	0.001	-"-
-"-		eq. w.	0.001	EIC 80
Argillite	Eleana	g.w.	0.17-0.83	DAN 81:2
-"-	-"-	oxic g.w.	0.011	ERD 79
-"-		-"-	<0.040	ERD 78
-"-	Eleana	-"-	0.025	ERD 79:8
-"-	Syncline Ridge	synth. g.w.	0.016-0.196	LYN 79
-"-	Calio Hills	-"-	0.155-6.5	-"-
Tuff	devitrified	oxic g.w.	0.0003	WOL 80
-"-	zeolitized	-"-	0.0002	-"-
-"-	devitrified	g.w., N_2 atm.	0.008-0.026	-"-
-"-	zeolitized	-"-	0.013	-"-
-"-	Yucca Mountain	-"-	0-013-0.026	CRO 80:2
-"-		oxic g.w.	<0.001	VIN 80
-"-	LASL	preeq. w.	<0.001	STR 80
-"-	NTS	-"-	<0.001	-"-
-"-		basalt w.	0	TEW 78
Silt	Muscatine silt loam	0.01 meq. Ca/l	0.00009	FRA 77
-"-	mottled siltstone		<0.001	STR 80
Shale	Conasauga	limestone w.	0.0001	TEW 78
-"-	green argillaceous	-"-	0	-"-
-"-	Pierre	peeeq. w.	<0.001	STR 80
-"-	PNL	-"-	<0.001	-"-
-"-		eq. w.	0.001	EIC 80
Hornfels		synth. g.w.	0	LYN 79
Limestone	PNL	preeq. w.	<0.001	STR 80
-"-		eq. w.	0.001	EIC 80
-"-	Cherry canyon	g.w. and brine	0.002-0.007	SER 77

Table V 2(3)

Rock	Origin/type	Water	K_d m^3/kg	Reference
Dolomite	synthetic	synth. g.w.	0	DOS 79
-"-	Culebra	-"-	0.001	-"-
-"-	Magenta	-"-	0.015	-"-
-"-	NTS	preeq. w.	<0.001	STR 80
-"-	Culebra	g.w. and brine	0.0002-0.0005	SER 77
Mineral group	Mineral			
Cyclosilicates	chrysocolla	preeq. w.	0.01-0.02	STR 80
Inosilicates	augite	0.03 N	0	REL 77
-"-	hornblende	-"-	0	-"-
-"-	enstatite	-"-	0	-"-
Phyllosilicates	biotite	granite w.	0	TEW 78
-"-	-"-	0.03 N	0	REL 77
-"-	kaolinite	-"-	0	-"-
Tectosilicates	quartz	-"-	0	-"-
-"-	albite	-"-	0	-"-
-"-	anorthite	-"-	0	-"-
-"-	microcline	-"-	0	-"-
Oxides/hydroxides	Al_2O_3	0.01 M NaCl	0.005	MEY 80:2
-"-	ZrO_2	-"-	0.03	MEY 80
-"-	bauxite	synth. g.w.	0	RAN 79
-"-	laterite	-"-	0	-"-
-"-	limonite	-"-	0.006	-"-
-"-	goethite	-"-	0.009	-"-
Carbonates	siderite	-"-	0.060	-"-
-"-	witherite	preeq. w.	<0.001	STR 80
-"-	cereusite	synth. g.w.	0.002	RAN 79
Phosphates	apatite	preeq. w.	<0.001	STR 80
Sulfates	anhydrite	-"-	<0.001	-"-
Sulfides	pyrite	-"-	<0.003	-"-
-"-	chalcopyrite	-"-	<0.001	-"-
-"-	-"-	synth. g.w.	0	RAN 79
-"-	galena	preeq. w.	0.001-0.010	STR 80
-"-	-"-	synth. g.w.	0.040-3.0	RAN 79
-"-	bornite	preeq. w.	<0.003	STR 80
-"-	CdS	-"-	0.00003	-"-
-"-	Cu_2S	-"-	0.00099	-"-
-"-	CuS	-"-	0.00004	-"-
-"-	FeS	-"-	0.00099	-"-

Table V 3(3)

Mineral group	Mineral	Water	K_d m^3/kg	Reference
Sulfides	HgS	preeq. w.	0.00002	STR 80
-"-	PbS	-"-	0.00046	-"-
-"-	ZnS	-"-	0.00008	-"-
Sulfosalts	tetrahedrite	tuff g.w.	0.028	ERD 81:2
-"-	-"-	preeq. w.	0.03-0.016	STR 80
-"-	bournonite	tuff g.w.	0.10	ERD 81:2
-"-	-"-	preeq. w.	0.2-2.0	STR 80
-"-	enargite	-"-	<0.001	-"-
-"-	tennantite	-"-	<0.003	-"-
-"-	boulangerite	tuff g.w.	0.088	ERD 81:2
Clay				
Montmorillonite		0.03 N	0	REL 77
Vermiculite		-"-	0	-"-
Illite		-"-	0	-"-
Hectorite		5 N NaCl	0.001	NOW 80
Soil				
Abyssal red clay		0.68 N NaCl	0	ERI 79:2
Fuqay sand		0.01 meq. Ca/l	0.00002	FRI 77
LT metamorphic		preeq. w.	<0.001	STR 80
South Carolina subsoil		0.03 N	0	REL 77

TABLE VI. Distribution coefficients for iodine. Low and medium ionic strength, pH 7-9.

Rock	Origin/type	Water	K_d m^3/kg	Reference
Granite		preeq. w.	0.001	FRI 78
-"-		-"-	<0.001 (I^-)	STR 80
-"-		-"-	<0.001 (IO_3^-)	-"-
-"-	PNL	-"-	<0.001 (IO_3^-)	-"-
Basalt	Sentenniel Gap	0.01 M NaCl	0	MEY 79
-"-		preeq. w.	0.002	FRI 78
-"-	CR	-"-	0.002(I^-)	STR 80
-"-	-"-	-"-	<0.001 (IO_3^-)	-"-
-"-	PNL	-"-	<0.001 (IO_3^-)	-"-
Tuff	devitr., zeolitized	equil. w.	0	JOH 80
-"-	Rainier Mesa		0.001	BOR 76
-"-	Los Alamos	preequil. w.	0.0008	FRI 78
-"-	NTS	-"-	0.001	-"-
-"-	LASL	-"-	<0.001 (I^-)	STR 80
-"-	-"-	-"-	<0.001 (IO_3^-)	-"-
-"-	NTS	-"-	<0.001 (I^-)	-"-
-"-	-"-	-"-	<0.001 (IO_3^-)	-"-
Alluvium			0.020	WOL 78:2
Siltstone		-"-	0.005	FRI 78
Shale		-"-	0.002	-"-
-"-	Pierre	-"-	<0.001 (I^-)	STR 80
-"-	-"-	-"-	0.001 (IO_3^-)	-"-
-"-	PNL	-"-	0.009 (IO_3^-)	-"-
Limestone	oolitic	0.01 M NaCl	0	MEY 79
-"-	PNL	preequil. w.	0.016 (IO_3^-)	STR 80
Dolomite	Culebra	0.01 M NaCl	0	MEY 79
-"-	Magenta	-"-	0	-"
-"-		g.w.	<0.001	DOS 78
-"-		preequil. w.	0	FRI 78
-"-	NTS	-"-	<0.001 (I^-)	STR 80
-"-	-"-	-"-	<0.001 (IO_3^-)	-"-
Chalk		-"-	0.005	FRI 78
-"-		-"-	<0.001 (I^-)	STR 80
-"-		-"-	0.005 (IO_3^-)	-"-

Table VI 2(4)

Mineral group	Mineral	Water	K_d m^3/kg	REference
Cyclosilicates	chrysocolla	g.w.	0.010	RAN 79
-"-	-"-	preequil w.	0.063	FRI 78
-"-	-"-	-"-	0.003-0.012 (I^-)	STR 80
-"-	-"-	-"-	0.04-0.075(IO_3^-)	-"-
Inosilicates	augite	0.01 M NaCl	0	MEY 79
-"-	-"-	0.03 N	0	REL 77
-"-	hornblende	0.01 M NaCl	0	MEY 79
-"-	-"-	0.03 N	0	REL 77
-"-	enstatite	-"-	0	-"-
Phyllosilicates	biotite	-"-	0	-"-
-"-	serpentine	0.1 M NaCl	0.001-0.002	MEY 79
-"-	kaolinite	0.03 N	0	REL 77
Tectocsilicates	quartz	-"-	0	-"-
-"-	-"-		0	ROU 79
-"-	albite	0.03 N	0	REL 77
-"-	anorthite	-"-	0	-"-
-"-	microcline	-"-	0	-"-
Oxides/hydroxides	Al_2O_3	preequil. w.	0	FRI 78
-"-	CuO	-"-	0	-"-
-"-	bauxite	synth g.w.	0.001	RAN 79
-"-	limonite	-"-	0.006	-"-
-"-	goethite	-"-	0.001	-"-
-"-	laterite	-"-	0.0002	-"-
-"-	PbO_2	preequil. w.	0	FRI 78
-"-	Fe_2O_3	-"-	0.005	-"-
Carbonates	azurite $Cu_3(CO_3)_2(OH)_2$	-"-	0.052-0.108	-"-
-"-	siderite $FeCO_3$	synth. g.w.	0.0002	RAN 79
-"-	cerussite $PbCO_3$	-"-	0.003-0.008	-"-
Phosphates	apatite	preequil. w.	0.0008-0.002	FRI 78
Sulfates	anhydrite	0.01 M NaCl	0	MEY 79
-"-	-"-	preequil. w.	0.005	FRI 78
Sulfides	pyrite	-"-	>0.330	-"-
-"-	-"-	-"-	0.3-3.0 (I^-)	STR 80
-"-	-"-	-"-	>1.0 (IO_3^-)	-"-
-"-	galena	0.01 M NaCl	0.001-0.013	MEY 79
-"-	-"-	preequil. w.	0.052	FRI 78
-"-	-"-	g.w.	0.015-0.050	RAN 79
-"-	-"-	preequil. w.	0.026 (I^-)	STR 80
-"-	-"-	-"-	0.003-0.045(IO_3^-)	-"-

Table VI 3(4)

Mineral group	Mineral	Water	K_d m^3/kg	Reference
Sulfides	CuS	preequil. w.	>0.330	FRI 78
-"-	Cu_2S	-"-	0.330	-"-
-"-	chalcopyrite	-"-	>0.330	-"-
-"-	-"-	g.w.	0.215	RAN 79
-"-	-"-	preequil. w.	0.190 (I^-)	STR 80
-"-	-"-	-"-	0.400-1.30(IO_3^-)	-"-
-"-	bornite Cu_5FeS_4	-"-	0.163	FRI 78
-"-	-"-	-"-	0.108-0.163	-"-
-"-	-"-	-"-	0.030-0.970 (I^-)	STR 80
-"-	-"-	-"-	0.01-0.10 (IO_3^-)	-"-
-"-	CdS	-"-	0.107	FRI 78
-"-	HgS	-"-	0.008	-"-
-"-	PbS	-"-	0.108	-"-
-"-	ZnS	-"-	0	-"-
Sulfosalts	tetrahedrite$Cu_{12}Sb_4S_{13}$	-"-	>0.330	-"-
-"-	-"-	tuff g.w.	0	ERD 81:2
-"-	-"-	preequil. w.	1.1-1.5 (I^-)	STR 80
-"-	-"-	-"-	1.0-3.0 (IO_3^-)	-"-
-"-	-"- (contains Ag)	-"-	0.004-0.68 (I^-)	-"-
-"-	-"- -"-	-"-	0.032 (IO_3^-)	-"-
-"-	bournonite $PbCuSbS_3$	-"-	>0.330	FRI 78
-"-	-"-	tuff g.w.	0.140	ERD 81:2
-"-	-"-	preequil. w.	0.200-1.0 (I^-)	STR 80
-"-	-"-	-"-	0.960 (IO_3^-)	-"-
-"-	enargite Cu_3AsS_4	-"-	>0.330	FRI 78
-"-	-"-	-"-	0.13-0.15 (I^-)	STR 80
-"-	-"-	-"-	0.07-0.12 (IO_3^-)	-"-
-"-	tennantite $Cu_{12}As_4S_{13}$	-"-	0.108	FRI 78
-"-	-"-	-"-	0.1-0.35 (I^-)	STR 80
-"-	-"-	-"-	0.01-0.02 (IO_3^-)	-"-
-"-	boulangerite $Pb_5Sb_4S_{11}$	tuff g.w.	0.140	ERD 81:2
Clay				
Montmorillonite		0.01 M NaCl	0	MEY 79
-"-		0.03 N	0	REL 77
-"-		dist. w.	0.001	FUR 73
Vermiculite		0.03 N	0	REL 77
-"-			0.001	ROU 79
Illite		0.03 N	0	REL 77
-"-			0.001	ROU 79

Table VI 4(4)

Soil	Water	K_d m^3/kg	Reference
Clay from a halite core	g.w.	0-0.004	DOS 78:2
Conasauga shale	tap w.	0.031-0.071	BRO 53
-"-	0.01 M NaCl	0	MEY 79
Green argillaceous shale	-"-	0	-"-
Muscatine silt loam		0	FRA 77
Oak Ridge clay	tap w.	<0.0001	LAC 54
Boom clay		0.001-0.005	DER 79
Glauconite sand		0.001-0.005	-"-
Diatomite	dist. w.	0.002-0.003	FUR 73
Silicagel	-"-	0.001	-"-

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