CHAPTER - VI

CHAPTER 6

GENESIS OF BENTONITE DEPOSITS

Background information

Bentonite is a secondary product. It is mostly derived from parent material of volcanic origin such as basalt, obsidian, rhyolite, volcanic tuff and ash. These parent materials contain in many cases, a considerable amount of glass, which is more vulnerable to hydrothermal and weathering activity than the crystallised silicates and commonly give rise to major formations of bentonite.

Bentonites form as a result of weathering of volcanic ash and rocks. Keller (1957) has shown that at alkaline pH and in a closed environment, volcanic material can evolve into zeolites, especially analcime or into montmorillonite.

When tholeiitic basalts, which are basic crystalline volcanic rocks having olivine, pyroxene and calcic plagioclases as the abundant minerals, are subjected to weathering in a temperate to tropical climate, the silicate minerals are broken down and depending on the pH of the environment, montmorillonite or kaolinite is formed. Montmorillonite is formed in an alkaline environment.

Millot, Bonifas, (1955), have shown that bentonite deposits can be formed due to hydrothermal action on volcanic material.

In the present study area the consanginious occurrence of the bentonite and laterite has compelled the author to envisage a probable genetic relationship between the two. The formation of bentonite has been recognised as a phase of the major process of lateritisation. The author has briefly pointed out the various theories of laterite formation.

The theory that laterite was the residuum of extreme weathering and differential removal of material by chemical solution was widely accepted. Glinka (1914) reinforced and elaborated Russel's (1889) theories of tropical weathering and formation of laterite residues.

The concept of laterite as a residuum has persisted, often supported by convincing evidence (deVletter, 1955; Hartman, 1955; du Bois and Jeffery, 1955; Hanlon, 1945; Van Bemmelen, 1941), but from about the first decade of the new century there was generally, progressively less stress put upon the residual aspect and more upon the precipitationary aspect.

The concept of laterite as a precipitate resulted from recognition of the fact that pallid zones were thought to be typically underlying the sub-strata laterites. This coincided with an increasing understanding of the several ways in which iron and alumina can be mobilised. Thus it came to be believed that the enrichment was caused by the iron and alumina, in solution, moving into the enriched zones and then being precipitated. The provenance of concentration was believed to be essentially the underlying pallid zone, but more distant sources were also considered possible if not probable (in Mc Farlane, 1976, p 93).

The role of groundwaters in the formation of laterite was first suggested by Maclaren (1906), who stated that laterite is not the result of decomposition of in-situ material. The ground waters may have derived their mineral contents from the underlying rocks but may also have brought it from sources far distant. With this concept, the stress on the immobility of the precipitates as the means of the accumulation lost ground to the suggestion that the accumulations were in large part due to precipitation from groundwaters, and the source of the enrichment of these groumdwaters was below the laterite but not necessarily local (in Mc Farlane, 1976, p.4).

Campbell (1917) was the first to formulate definitely the concept of laterite as a precipitate. Since its precipitationary aspect was already well recognised, he must be understood to have meant that the accumulated minerals had much greater degree of mobility than was formerly recognised, i.e., the concentration was due to this mobility. This was a major turning point in the development of theories of laterite genesis and was responsible for bringing the material into the sphere of interest of pedologists.

Two mechanisms were suggested for this enrichment of the laterite horizon; capillarity and the seasonal fluctuation of the water table. Laterite formation by capillary action was favoured by Mc Farlane (1906). The evidence against the effectiveness of capillary action in laterite formation is overwhelming (Goudie, 1973, p. 141-144). Not least is the recognition that capillary action is restricted to a very narrow horizon in the soil (Sivaraja-Singham et.al. 1962; Baver, 1956) seldom exceeding 2 m. Capillarity as a major factor, has generally been discredited but not entirely abandoned as a mecahanism by which some duricrusts are believed to accumulate (Loughnan et.al. 1962; American Geological Institute, 1962; Stamp, 1961).

The belief that upward movement of iron and alumina was brought about by water table fluctuations was more popular (Pendleton and Sharasuvana, 1946; Pendleton, 1941; Prescot & Pendleton, 1952; Fisher, 1958) and it is still very widely held. Essentially the theory is that enriched solutions are carried upward in the profile with the seasonal rise of the water table and precipitated near the upper limits of the range of fluctuation.

Trendall's (1962) theory of apparent peneplanation also regards laterite as a sort of residual precipitate. The main source of enrichment was visualised as being the pallid zone. The entire profile was believed to move downwards as the surface is lowered by wash, and thus the zone of water table fluctuation progressively incorporates new material from which it draws iron for the laterite horizon.

A further theory has been evolved in Uganda (Mc Farlane, 1971). It also regards laterite as a residual precipitate. The ground water laterite is believed to accumulate as a mechanical residuum during the late stages of reduction of downwasting land surface. The original precipitates form within the relatively narrow range of fluctuation of the ground water table, which sinks as the land-surface is reduced by erosion. They become incorporated into the lower parts of the soil mantle where they accumulate as an increasingly thick layer. When down wasting has ceased and the water table has stabilised, the residuum is hydrated and altered, and a massive variety of laterite develops which has every appearance of a true precipitate. This alteration of a residuum or detrital laterite is very similar to the so called detrital model, but the vertical lowering of the low-relief land surface allows the detritus to cover thickly all parts of it. Du Bois and Jeffery (1955) suggested that not merely part, but all of the enrichment is derived from an overhead source, here also no contribution is believed to have been provided by the pallid zone which underlie the high level laterites. This theory is in part based on the study of the low-level laterites which can be seen to be at various stages of this sequence of land surface reduction and laterite development (Mc Farlane, 1969).

Ida Valeton (1983) described the formation of lateritic bauxites world-wide in Cretaceous and Tertiary coastal plains. The bauxites form elongated belts, sometimes hundreds of kms long, parallel to lower Tertiary shorelines in India and South America and their distribution is not related to a particular mineralogical composition of the parent rock. The lateral movement of the major elements Al, Fe, Ti is regarded by Valeton as dependent on a high level flow of groundwater. Varying efficiency of sub-surface drainage is believed to produce lateral facies variation. Lateritic bauxites are considered to be a part of an alteration blanket, which is formed by in-situ pedogenic processes leading to extremely intensive geochemical separation of Si, Al, Ti and Fe. This process always leads to a vertical division into three major soil horizons.

Horizon rich in oxides	B ox, Fe, Al
Horizon rich in silicates (saprolite)	B sap
Horizon of fresh parent rock	C

Normally the soil sections are truncated. The A-horizon is always eroded (Valeton, 1983). After the Soil Taxonomy of the U.S. Department of Agriculture (1965), this type of lateritic bauxite belongs to the sub-group of aquox in the group of oxisols which are soft during the time of formation. During uplift above the groundwater level, the Fe-rich parts form the hard ferricretes wheras the Al-rich parts become hard alucretes (Goudie, 1973).

Lateritisation Process

Lateritisation is a process of weathering and can be described as the formation of a new mineral phase equillibrium (Mc Farlane, 1976). He stated that most rocks outcropping at the earth's surface are in a mineral phase equillibrium related to the higher temperature and pressure than is found at the surface. During the process of lateritisation the primary mineral in the parent rock are broken down, some components are leached out and secondary minerals form from the residue. These inturn are broken down and suffer the loss of some of their components such that the reidue regroup to form more stable secondary minerals in a continuing process. The mineral progression is therefore generally towards greater stability in situations progressively nearer the surface. Laterites are the product of intense sub-aerial rock weathering whose Fe content is higher and Si content lower than in merely kaolinised parent rock (Schellman, 1981). The saprolite horizon in a lateritic weathering profile is considered to be a half way station in the re-equillibration process of lateritic weathering and the bauxite formation as the extreme stage of lateritisation (Bardossy & Aleva, 1990).

In order to achieve such a selection several physical and chemical factors come into play and they are:

a) Climatic conditions

b) Topography

c) pH of the hydrolysing solutions

d) Nature and composition of the parent rock.

Climatic conditions

When a strongly leaching condition occurs in a tropical climate, lateritisation takes place (Caroll 1962).

The formation of laterite is favoured by humid tropical climates with a long dry season that allow successive small scale distribution of iron by leaching, affecting the upper part of weathering profile (Boulet, 1978).

Wet and dry seasons are considered to be favourable, if not essential, to laterite genesis (du Preez, 1954; Sabot, 1954; Fisher, 1958).

Chorley (1975) has stated that under alkaline conditions, Al_2O_3 is very soluble and SiO_2 is relatively soluble, so both are carried in solutions, provided enough rainfall is available. If the rainfall is scanty, the decay products Al_2O_3 and SiO_2 are not removed but combine to form clay minerals such as montmorillonite and illite.

Montmorillonite forms under alkaline and poorly leaching conditions and in presence of Ca and Mg ions and in deficiency of K ions (Bates, 1952).

Climate is a paramount factor in chemical weathering. Rainfall controls the supply of moisture for chemical reactions and for the removal of soluble constituents of the minerals, while temperature considerably influences the rate of these reactions. According to "Van't Hoff's rule" for each $10^{\circ}C$ rise in the temperature the velocity of chemical reaction increases by a factor of from 2 to 3.

The climatic conditions of laterite formation are currently satisfied only between the latitude 30 and 30 S which coincides with Koppen's "A" climate zone of equitorial rainy, tropical monsoon and tropical wet and dry climate (Selby, 1971).

In order to explain the formation of Bhavnagar laterites/bentonites along with thick forests present at that time, the author has taken into consideration the drift tectonics of the Indian plate as it was moving across the equatorial zone. As shown in fig 24 , the Bhavnagar laterites, when plotted on a map of India along with other Palaeocene and Eocene laterite/bauxites of Gujarat, Rajasthan, NW Uttar Pradesh and Jammu-Kashmir states of India, show a very linear alignment (Sychanthavong & Patel, 1987). This linear belt is now trending in a NE-SW direction. If we consider the palaeoposition of the Indian plate during its early stages of drifting from the Gondwana land, it is found that it was trending E-W coinciding with the alignment of the latitudinal positions (Fig 25). It can be implied that this Palaeocene laterites must have been developed when the Indian plate was in the equatorial zone. The time span for the development of these laterites was quite short. This corresponds to the speed and the time spent by the Indian plate in the equaitorial zone as determined by the studies of the magnetic anomalies of the Indian Ocean and Arabian Sea floor (Sychanthavong & Patel, 1987). Thick tropical forests of this

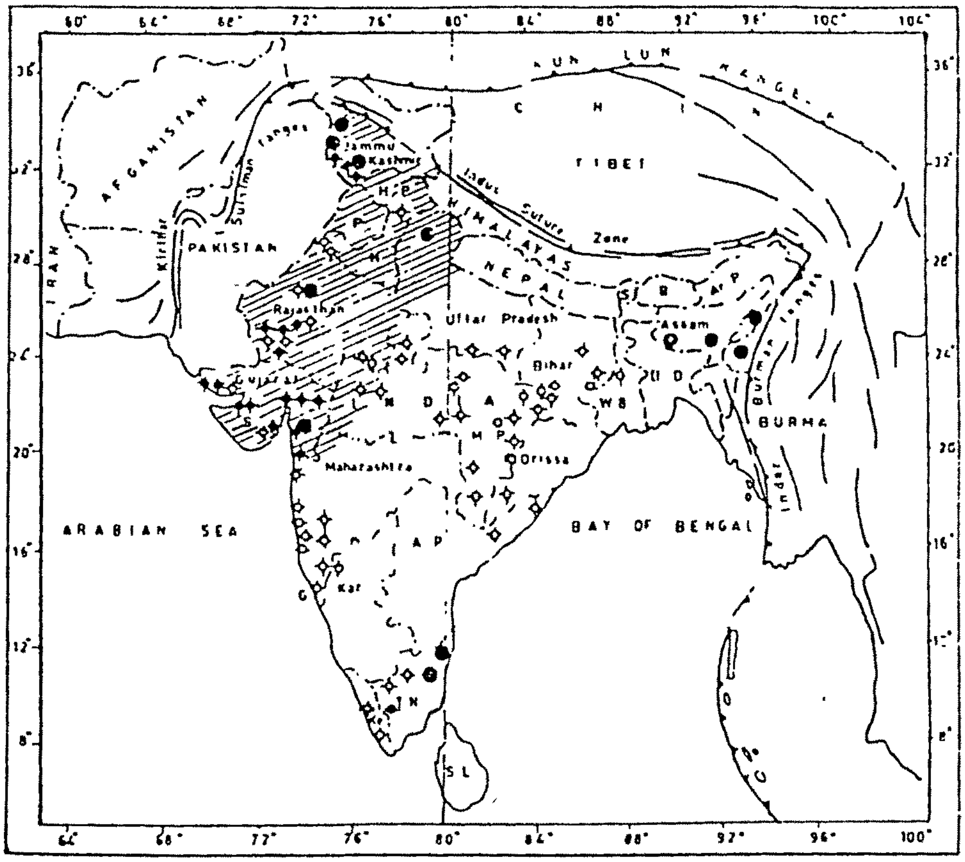


Fig 24: Distribution of Tertiary laterites, bentonites, bauxites and lignites (with gradual decrease in age from north to south). The plots within stripped lined area are Palaeocene in age. Filled circles with four sided dashes are also laterites and bauxites; open circles with two sided dashes denote bentonites and Fuller's earths; and solid circles are lignites and coals. K and S in Gujarat state indicate Kutch and Saurashtra respectively; and the other alphabetical symbols correspond to the names of various Indian states.
(after Sychanthavong and Patel 1987)

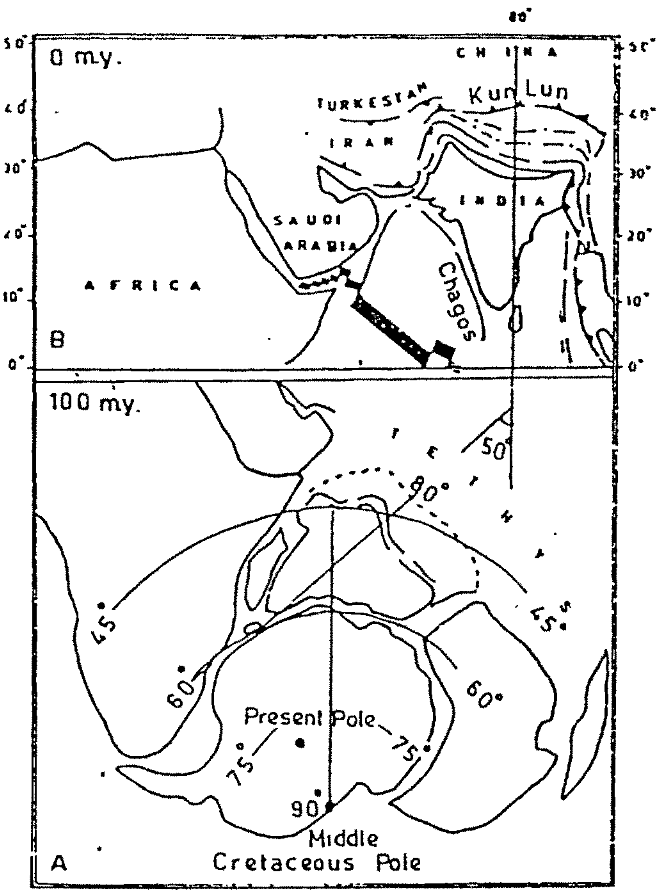


Fig 25: A – The palaeopositions of the Indian plate and other sorrounding continents in the Gondwanaland (after the best fit of Smith and Hallman, 1970). This illustrates the position of the Indian plate with tho alignmcnts of the then latitudes and longitudes. The location of the Middle Cretaceous pole is after Creer (Earth Science Reviews, 6, 369-446). The shape of India is presented in terms of Greater India's Concept, indicated by broken line. B – Present configuration of the Indian plate having been rotated anticlockwise for a total amount of 50 with respect to its original position in the Gondwanaland.
(after Sychanthavong and Patel, 1987)

period would not have survived under a climate with lesser rainfall which was encountered when the northern part of the Indian plate moved above the equator prior to 53 million years. These forests were ultimately destroyed and deposited as lignite seams in depressed areas during the early Eocene. Hence it is logical to suggest that the Indian plate was rotated by 40° in an anticlockwise direction, after this particular linear laterite/bauxite belt had crossed the equitorial zone while the total rotation was 50° (Fig 26). Thus the Indian plate was rotated by only 10° or even les during its journey in the southern hemisphere. This geological evidence is in agreement with the reconstruction of the palaeo position of the Indian plate and the age of the ocean floors, right from the early stage of its breakup and the opening of the Indian Ocean and the Arabian Sea (Sclater & Fisher, 1974). According to this reconstruction, it is found that southern India spent a longer time in the equatorial zone, i.e. between 53 m.y. and less than 32 m.y.. It has been discovered that the initial speed of drifting of the Indian plate was very fast, 17 cm/yr total rate, it decelerated and then stopped around 56 m.y. ago, and again started drifting with a much slower rate (McKenzie & Sclater, 1971). This has been possibly attributed to its leading edge having reached the subduction zone in the Tethys sea to the north. This must be the only logical reason to explain why the southern Indian continental plate was spending a lot of time in the equatorial zone (Sychanthavong & Patel, 1987).

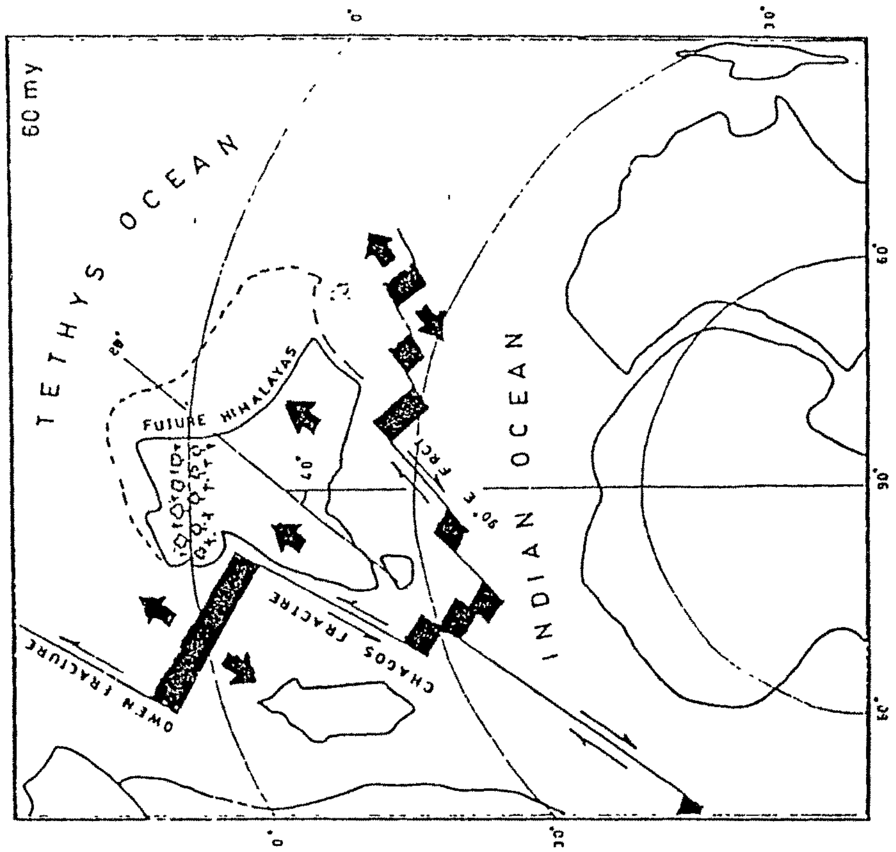


Fig 26: The palaeoposition of the Indian plate during which the northwestern laterite and lignite belt was being formed, indicated by forested area. Broken line indicates the possible extention of Greater India which we believe to have drifted away much earlier than did the main Indian plate.
(after Sychanthavong and Patel, 1987)

Frakes and Kemps (1972) completed the continent reconstructions by ocean current evaluation, by sea water temperature determinations based on oxygen isotope measurements and by palaeo-botanic data about the land vegetation. The resulting palaeoclimatic map shows that during the Eocene (Fig 27) the conditions were extremely favourable allowing laterite/bauxite formation upto 50° - 53° latitudes.

Valeton (1972) stated that in Eocene times, formation of bauxite had wide spread golbal geographical range.

A pointer to the palaeoclimatic conditions in the study area are the laterite cappings which overlie deep weathered residual profiles. This indicates that the moist climate was followed by an alternating wet and dry season during the period of lateritization (Woolnough, 1918).

Also, the colour banding which is present in the bentonite zone is due to the alternating enrichment and depletion, mostly of the iron oxides, resulting from the patchy and periodic drying up of the ground water (Ollier, 1969). This can only occur where there is an alternating wet and dry season.

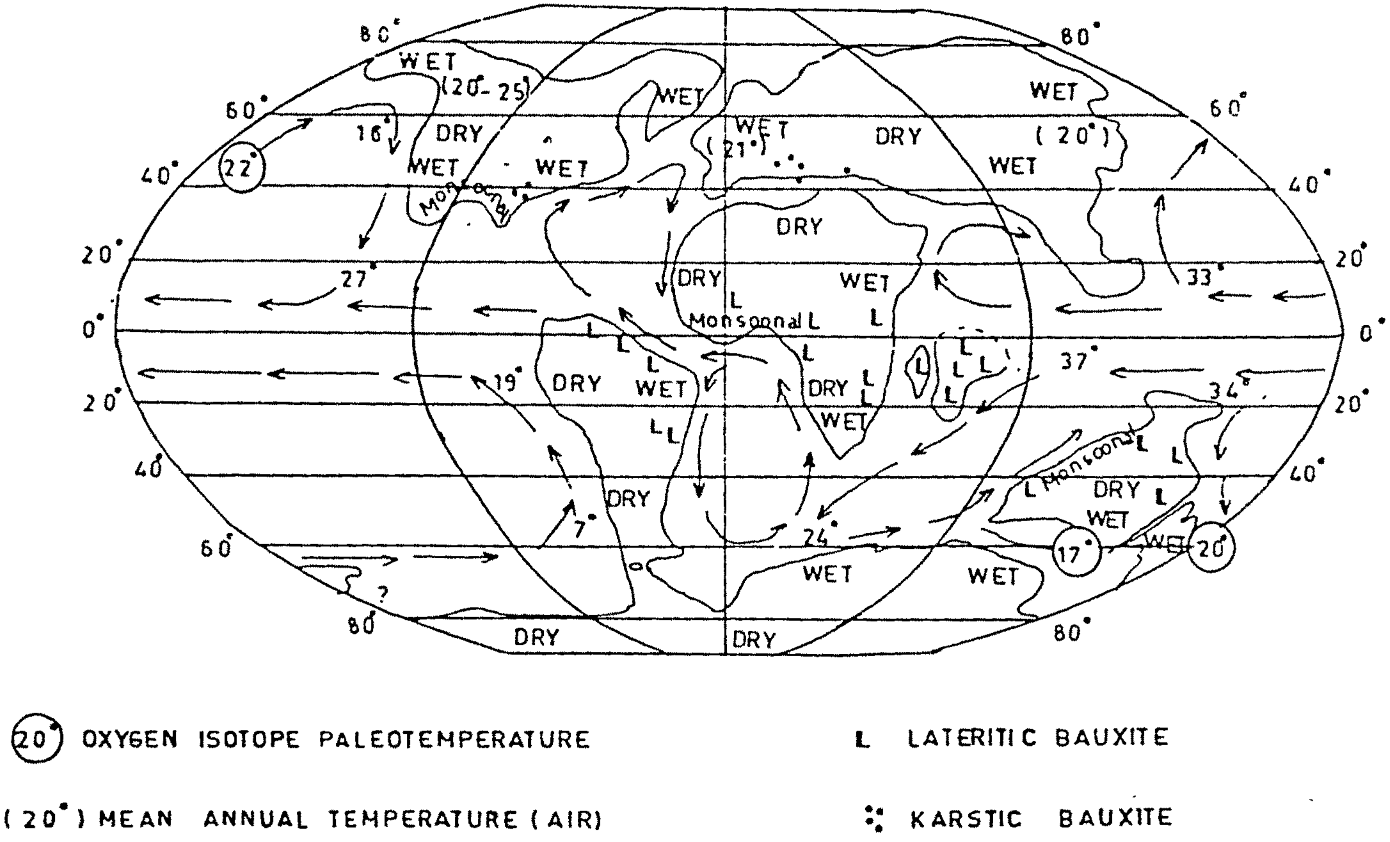
Hence, the climate during the period of lateritisation must have been tropical to sub-tropical with alternating wet and dry seasons favouring the formation of bentonites.

Topography

Topography has a marked effect of the rate of chemical weathering and on the nature of the weathered products. It exerts this

FIG 27 : RECONSTRUCTION OF WEATHER PATTERNS IN EOCENE

(After Frakes and Kemp, 1972)



influence in several ways by controlling (a) the rate of surface runoff of rainwater and hence the rate of moisture intake by the parent rock, (b) the rate of subsurface drainage and therefore the rate of leaching of the soluble constituents and (c) the rate of erosion of the weathered products and thereby the rate of exposure of fresh mineral surfaces.

To achieve efficient leaching and solute removal, particularly good drainage is needed; high permeability of the parent rock being an essential condition for this (Bardossy & Aleva, 1990).

Valeton (1972) states that for the solution, migration and precipitation of major and minor elements within the bauxite alteration blanket, the following ground water conditions must be fulfilled;

a) net flow towards the sea,

b) groundwater levels must be high and oscillatory nature and

c) the environment must be reducing.

A moderate to gently sloping surface allows free movement of water table or sub-surface drainage, with a minimum of erosion and also helps to diminish loss of water by run-off (Esgura, 1964).

If topography is low, then the level of ground water is high and length of vertical percolation is small, therefore weathering of rocks is slight. Steep slopes as in a rugged topography cause rapid run-off and indiscriminate removal of weathering products.

Moss (1965) noted that while mottled ferallitic clays are rarely associated with slopes of more than 4°, the non-mottled ferallitic clays are found associated with slopes of 1° or less.

Woolnough (1918) noted that "lateritisation can occur only in areas where the drainage is almost at a stand still".

Altschuler et at (1963) opined that montmorillonitisation occurs below the ground water table level in sluggish drainage conditions. Above the water table level, the montmorillonite starts decomposing into kaolinite due to freer drainage conditions.

From the bore hole records of the Directorate of Geology and mining in the study area it has been inferred that the basalts have a very gentle dip towards the coast.

pH of the Hydrolysing solution

Chemical weathering is caused by the hydrolysis of rock minerals. Water with its content of cations and anions is the reagent that causes hydrolysis; the amount of H dissociated in water is indicated by the pH of water, which is the expression of the chemical environment of weathering (Caroll, 1962).

The pH of natural water is controlled by chemical reactions and equillibrium among the ions in solution in it. The pH of rain water is primarily due to the absorption of CO_2, and at 25° C when it is saturated with CO_2, the pH is approximately 5. Rain water in clouds has a pH of 5.7, but this value varies according to the other cations that are present (Caroll, 1962). Not all rainwater is saturated with CO ; there is a delicate equillibrium between the CO , H CO and HCO content which can be thermodynamically stated as

$$\frac{(H)^{+} \quad (HCO_3)^{-}}{H_2CO_3} = KH_2CO_4 = 10^{-6.4} \quad \text{at } 25^\circ C.$$

This constant varies from 6.58 at 0°C to 6.33 at 30°C (Garels and Christ, 1965, p.89)., which is the pH range in which weathering takes place.

The pH of mineral surface is determined by the abrasion pH - a term introduced by Stevens & Correns (1948). It is the pH that the mineral develops when pulverised under water. The feldspar yields a solution of pH 8-10, augite 8-11 and hornblende 10-11. The pH of the hydrolysing solution is very important for the transportation of alumina and silica in solution and their ultimate deposition. Under alkaline conditions (pH 10 or more) Al_2O_3 is very soluble and SiO_2 is relatively soluble (Chorley, 1975), so both are carried in solution, provided enough rainfall is available. If the rainfall is scanty, the decay products Al_2O_3 and SiO_2 are not removed but combine to form clay minerals such as illite and montmorillonite. Montmorillonite forming under alkaline and poor leaching conditions and in presence of Ca and Mg ions and in deficiency of K ion is also supported by Bates (1952).

Under neutral conditions (pH 7 to 8) Al_2O_3 is almost insoluble and SiO_2 is partly soluble, which is removed in solution if rainfall is heavy. The hydrated Al_2O_3 remains as residue and forms gibbsite.

As weathering processes develop, however, by reactions involving the free H^+ ion, the acidity of water in soil and rock will decrease due to chemical reaction.

Nature and Composition of the Parent rock

Basalt is found exposed as a base rock at the bottom of some quarry pits and river sections. The boundary between the basalts and the overlying bentonites is transitional.

Megascopically the rock is greenish black to black in colour with numerous amygdaloidal cavities pockmarking it. These cavities are filled up with secondary minerals like zeolite, secondary silica, etc. Shining metallic black specules, resembling magnetite are found pockmarking the whole rock. There are occassional dark red tinges on the rock.

Under the microscope the unweathered basalt is hemicrystalline in texture with a good amount of glassy material present.

Plagioclase feldspars form both big laths and tiny needles, the former occurring as phenocrysts or as randomly oriented plates showing ophitic and sub-ophitic textures, enclosed within the pyroxene. The pyroxene is dominantly augite and occurs either as granules in the groundmass or as phenocrysts. It also forms large crystals enclosing labradorite laths and gives rise to ophitic and sub-ophitic textures. Sporadic grains of olivine are seen embedded in the mafic ground mass of the basalt. It is considerably serpentinised or chloritised. Magnetite constitutes the major part of the iron ores, though some times hematite is also present. Both occur mainly in the ground mass.

Chlorites and calcite occur within the vesicular cavities. Chlorite also forms an important alteration product of olivine and augite. Zeolites too form secondary infilling material of vesicles and amygdales, forming fibrous colourless aggregates.

Laterite development has often been related to what are described as suitable rocks (Woolnough, 1918; Hanlon, 1945) but different authors have had conflicting opinions as to what constitutes suitability. Woolnough, for example, maintained that there is very little tendency to laterite formatiion on anything but the basic rock types (Woolnough, 1918, p.389). Simpson (1912), considered suitable rocks to be largely composed of metallic silicates, especially granite and greenstones of commom types, as well as amphibolites, epidiorites, chlorite schists and other similar rocks. Holmes (1914), noted that gneisses and basalts were particularly suitable, and Maud (1965) noted a preferential development on sandstone. Pallister (1954), observed a preference for metasediments, and Wayland (1935) for igneous rocks.

Certain rocks were considered unsuitable for laterite development. Holmes (1914), noted its absence on granites and sedimentaries. Maud (1965), also noted its absence and Pallister (1954) its poor development over granite.

In fact, it appears that laterite development is not confined to certain rocks, but over a great variety if not all rock types (McGee, 1880; du Preez, 1949; Mulcahy, 1961; Scrivenor, 1933; de Weisse, 1954; Walther, 1915; Gozan and Vera, 1962; Blondel, 1954; Campbell, 1917; Grubb, 18633; Mabbutt, 1961).

Depending upon the chemical and mineralogical composition of the parent rock, laterites are classified into the following groups : (Harder, 1951):

1) Those derived from rocks rich in alkali-aluminium silicates, particularly the felspathoids, with alkali felspars, pyroxenes, amphiboles, and micas, but no quartz:
2) Those closely associated with limestones and presumably resulting from the decay of their residual products:
3) Those derived from the weathering of hydrous aluminium silicate rocks such as sedimentary clays with little free quartz:
4) Those associated with intermediate and basic rocks such as diorite, gabbros, dolerite and basalt, containing abundant calcium-aluminium and calcium-magnesium-iron silicates, but little or no quartz:
5) Those resulting from the intensive weathering of various moderately aluminous igneous, metamorphic and sedimentary rocks such as granite, syenite, gneiss, schist, phyllite, slate and shale containing more or les quartz.

These anomalies are informative, for if in one area laterite development is favoured by a particular lithology while in another, its development is apparently inhibited by the same lithology. This suggests that other environmental factors are operative if not dominant in its development.

A classic example of this is illustrated by the laterites of Bhavnagar district. Here the occurrence is in the form of a narrow elongate belt, fringing the Deccan Trap basalts towards their Tertiary contact. From this occurrence itself it can be deduced that though the whole of the Trap basalts were experiencing the same climatic conditions, only a narrow strip of the basalt underwent lateritisation.

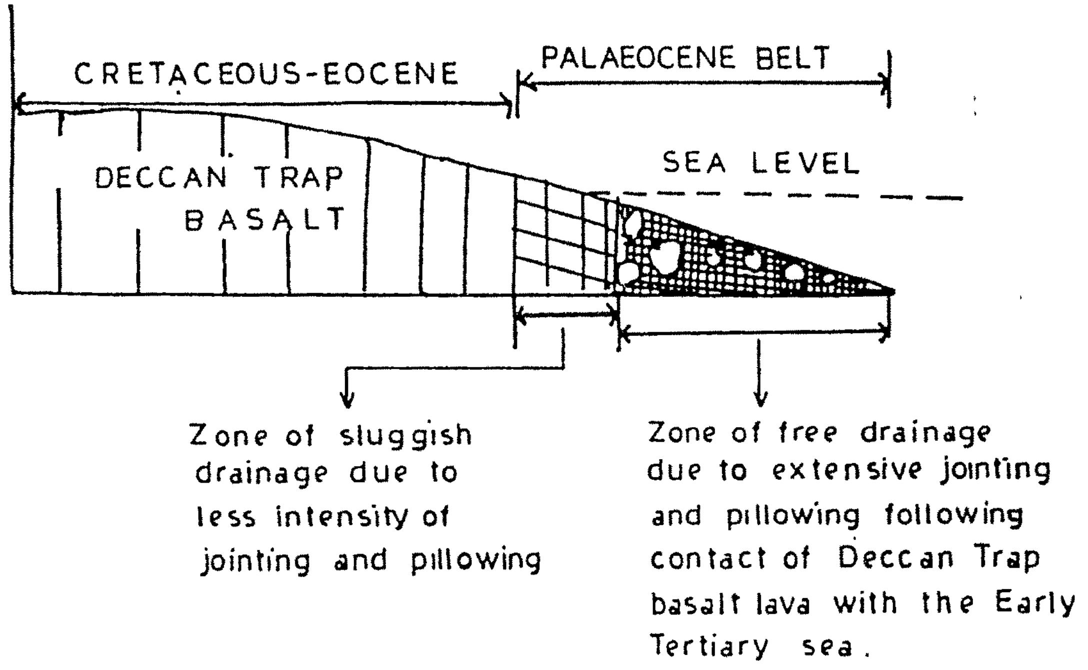
One can question as to why this narrow belt underwent lateritisation with the major portions of the Deccan Trap formation remaining unaffected. A possible cause can be that the linear early Eocene belt, representing the edge of the outpouring of the Deccan Trap basalts came in contact with the early Tertiary sea and this sudden cooling of the lava all along the palaeo-shoreline must have caused extensive pillowing and joint formation. This is manifested in the weathering profiles where extensive relict spheroidal structures are observed. These cooling structures paved the way for the easy access of rainwater and groundwater, which in turn accelerated the process of chemical decay of the parent Deccan Trap basalt along its leading edges (Fig 28).

Genetic Model for the bentonites of the Study area

In the study area the field characteristics and the residual relationship between the basalt, bentonite and the laterite point out to an in-situ weathering of the Deccan basalts. This has

FIG: 28

DIAGRAMMATIC REPRESENTATION OF THE NATURE OF THE LEADING EDGE OF THE DECCAN TRAP BASALT WHEN IT CAME IN CONTACT WITH THE EARLY TERTIARY SEA



resulted in the formation of each of the above mentioned rocks in close association within one profile making the bentonite occurrence quite unique.

The genetic model is based on the fact that the extremely well preserved relict textures in this Palaeocene(?) blanket indicates that the original volume of the parent rock was preserved, and that a series of weathering environment existed superimposed on one another with the residual products of each forming the parent material for the succeeding environment. The model is represented diagramatically in fig 29.

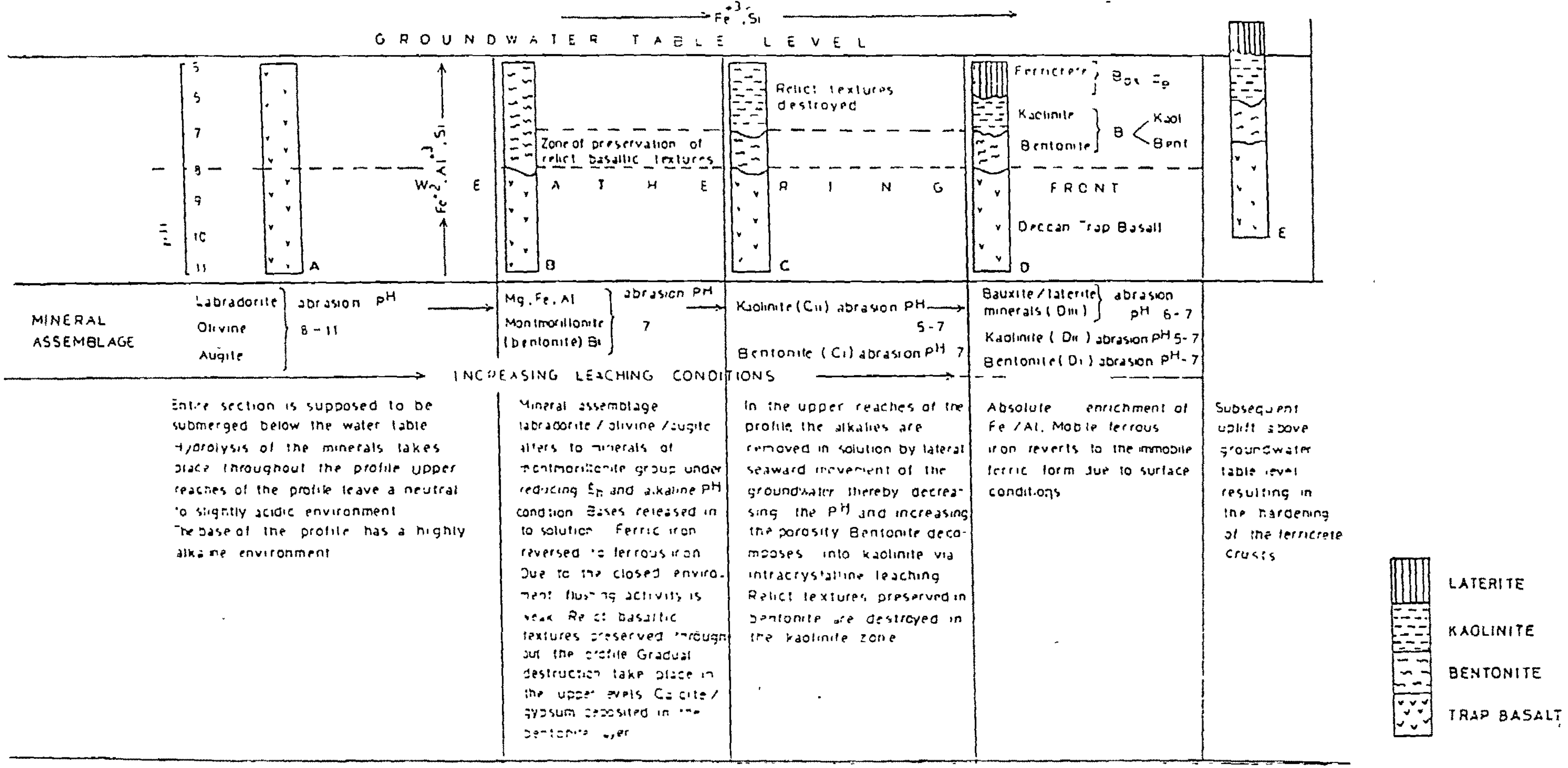
The main source of enrichment has been visualised as being the pallid zone. This is the zone underlying the laterite and is pale coloured in comparison to the laterite due to the leaching of iron. The pallid zone is synonymous to the belt of lithomarge and bentonite in Bhavnagar.

This model involves a large range of water table fluctuation. But in Bhavnagar, both the lithomarge and bentonite bands are in an extremely moist state indicating that there is a fair amount of water table fluctuation.

Formation of Bentonite

Under conditions discussed earlier the parent rock is attacked by rain water. The water percolates and comes in contact with the silicate rocks and the solutions of some rock forming compounds. The mineral lattices are destroyed and the co-ordination no longer controls the movement of cations.

FIG 29: MODEL FOR THE GENESIS OF LATERITE/BENTONITE IN BHAVANAGAR



	A	B	C	D	E
MINERAL ASSEMBLAGE	Labradorite Olivine Augite } abrasion pH 8 – 11	Mg, Fe, Al Montmorillonite (bentonite) Bi } abrasion pH 7	Kaolinite (Cii) abrasion pH 5-7 Bentonite (Ci) abrasion pH 7	Bauxite / laterite minerals (Diii) } abrasion pH 6-7 Kaolinite (Dii) abrasion pH 5-7 Bentonite (Di) abrasion pH-7	
	Entire section is supposed to be submerged below the water table Hydrolysis of the minerals takes place throughout the profile upper reaches of the profile leave a neutral to slightly acidic environment The base of the profile has a highly alkaline environment	Mineral assemblage labradorite / olivine / augite alters to minerals of montmorillonite group under reducing E_h and alkaline pH condition Bases released in to solution Ferric iron reversed to ferrous iron Due to the closed environment flushing activity is weak Relict basaltic textures preserved throughout the profile Gradual destruction take place in the upper levels Calcite / gypsum deposited in the bentonite layer	In the upper reaches of the profile the alkalies are removed in solution by lateral seaward movement of the groundwater thereby decreasing the pH and increasing the porosity Bentonite decomposes into kaolinite via intracrystalline leaching Relict textures preserved in bentonite are destroyed in the kaolinite zone	Absolute enrichment of Fe / Al. Mobile ferrous iron reverts to the immobile ferric form due to surface conditions	Subsequent uplift above groundwater table level resulting in the hardening of the ferricrete crusts

The initial belt which underwent lateritisation was as discussed earlier, highly jointed and with many pillow structures due to rapid cooling. This physical condition of the parent rock made it amenable to rapid chemical deterioration.

The most important leaching agent is water which acts in two ways
i) rain water percolating downwards and
ii) ground water fluctuating according to season

The ions Ca^{2+}, Mg^{2+}, K^{+}, Na^{+}, Al^{3+}, Si^{4+} Fe^{3+} which are released due to hydrolysis; majority of them are removed by percolating solution except Fe^{3+}, Si^{4+} and Al^{3+}.

Under most condition normally encountered in the ground water i.e., pH ranging from 3.5 to 8.0 , Al is less mobile than silica (Norton, 1973). With the change of rainy seasons to dry periods, the reaction of ground water also changes from slightly acidic to slightly alkaline. Most elements Ca, Na, Mg and K and part of silica are dissolved and migrate with ground water. Fe and Al are immobile until pH changes due to onset of next rainy season and vegetation growth. They move upwards and precipitate at the beginning of the next dry season (Neuzil & Kuzvart, 1964). Thus the weathering of the Deccan basalt at or near the ground water table, and the free movement of the latter cause removal of soluble elements and leave behind the mixture of Al hydrate with iron and some resistant minerals originally present in the parent rock.

The removal of silica occurs at higher pH conditions and such environment is quite often observed in soil and ground water (Bass Becking et al, 1976). Mohr (1933, 44) suggested that silica of minerals (in form of silicic acid) is easily soluble in rain water but water containing salts, carbonic acid or organic minerals decreases the solubility , so silica or kaolin moves in pure water while Fe oxides and alumina are insoluble in pure and salt solutions except in water containing organic material or strong acid.

Ferric iron is soluble in notable amounts if pH is less than 3. The iron is stable under strongly oxidising conditions and therefore immoblie (Craig & Loughnan, 1964). The presence of organic acids on the surface and a reducing environment locally developed in thickly vegetated areas probably causes iron to dissolve and form complex ions or colloidal hydroxides and migrate downwards. The required reducing environment may be enhanced by a supply of organic matter or an abundance of reducing minerals in the source rocks (eg. basalt) but it would be counter-acted by aeration (Peterson, 1971). Grim (1953) pointed out that due to high temperature in the tropics, organic matter is readily oxidised, so it does not accumulate. It seems therefore that protection by humus is unlikely to provide means by which large scale movement of iron can take place through aerated and neutral or alkaline horizons which could otherwise favour its stability.

Chemistry of the Weathering of Deccan Trap Basalts

The main minerals presents in the basalts are calcic plagioclase (labradorite), augite, olivine and magnetite.

Calcic Plagioclase Felspars

In the frame work structure of the felspars, the tetrahedra are linked through all four oxygen atoms to yield a three dimensional configuration. Aluminium substitution for silicon within the tetrahedra creates a net negative charge on the framework which is balanced by a calcium ion. Break down of the calcic felspars proceeds through the loss of these calcium ions, but for their escape, the tetrahedral framework must be ruptured. In the calcic felspars the aluminium- to-silicon ratio is 1:1. Despite the similarities in structure of the alkali and the calcic felspars, a considerable disparity exists in their weathering stability (Loughnan, 1969). Apparently, the greater replacement of silicon by aluminium in anorthite results in a higher degree of polarization of Al-O-Si bonds. With the loss of the metallic ions, the frame work structure breaks down into chains which tends to polymerize into sheets in much the same manner as those released from pyroxenes. Where the leaching conditions are inadequate to remove magnesium and ferrous ions as rapidly as they are released from the break down of associated minerals, these ions tend to be fixed by the new residual structures and montmorillonite results. Chapman & Greenfield (1949), Smith (1957) and Craig and Loughnan (1964), have described the alteration of calcic felspars to montmorillonite.

Olivine

Olivine is composed of discrete silica tetrahedra bonded together by magnesium and ferrous ions in an octahedral co-ordiation. Both octahedral cations are potentially mobile and their loss from the surface of the mineral causes the ready release of the individual tetrahedral unit, thereby exposing fresh surfaces to attack. Consequently olivine decomposes rapidly. However, the rate of release of the silica may exceed its rate of solution, in which case the residual silica polymerizes into sheets and apparently fixes some of the magnesia, yielding serpentine as the crystalline phase (Loughnan, 1969). Generally, a complete balance of silica and magnesia is attained so that neither free silica nor free magnesia remains within the weathering zone. This and the case with which the released silica can polymerize into the sheet structure of phyllocsilicates are phenomena which may be observed repeatedly in the weathering of the silicate minerals. Much of ferrous iron, if released above the water table, is oxidised, in which state it crystallises or tend to crystallise as either the oxide, hematite and more rarely maghemite, or hydroxide, goethite. However, much may persist in an extremely fine state and appear amorphous to X-ray radiation.

As weathering intensity increases, magnesia is preferentially leached form the serpentine which becomes progressively unstable and is converted into a new silica enriched phase, saponite (a montmorillonite group mineral).

It is not known whether complete destruction of the serpentine lattice is necessary for this transformation. Ultimately, saponite becomes unstable and both the remaining magnesia and silica are leached, leaving a residue of oxides and hydroxides of iron (Loughnan,1969).

Craig (1963), has shown that olivine may alter directly to magnesium-rich montmorillonite without passing through the serpentine stage, while Shermann, et al (1962) found nontronite the initial product of weathering of olivine.

Pyroxenes (Augite)

In the pyroxenes, the tetrahedra are arranged in chains which are bonded together by metallic ions, the most common being those which enter into octahedral co-ordination with oxygen, such as Mg^{2+}, Fe^{2+} , Fe^{3+}, Al^{3+}. Some alumina may replace silicon in the tetrahedra. Bonding by the octahedral cations is relatively weak and pronounced cleavage with planes approximately normal to each other parallel to the silica chains. Access of water by way of the cleavages promotes solution of the bonding cations and causes rapid breakdown of the structure. Upon release, the chains tend to polymerize into sheets incorporating residual alumina and magnesia forming chlorite or montmorillonite or both. Ferrous iron is stabilized by oxidation tothe ferric state, and at the same time titania crystallizes as anatase. Where the release of calcium from the breakdown of the pyroxene exceeds its rate of solution, calcite also develops.

As the intensity of weathering increases, all the lime and magnesia are lost together with the silica, not required to saturate alumina, and the residue becomes progressively enriched in kaolinite, anatase and oxides and hydroxides of ferric iron (Loughnan,1969).

Magnetite

This mineral oxidizes during weathering to yield hematite, magnetite and/or goethite.

It may be concluded that the initial weathering products of all the major mineral constituents of basalt are either nontronite or montmorillonite. This has been proved by the X-ray diffraction studies in the profiles at Badi, Thalsar, Morchand and Alang.

At this stage, the pore solutions are high in alkalies and alkaline earths derived from the decomposition of olivine and felspars. Due to this alkaline medium, gypsum and calcite are deposited in the bentonite belt. The ferric iron liberated is not mobile (Harden & Bateson, 1963; Pickering,1962) but can be mobilised by reversion to ferric iron under water logged conditions.

Upwards, the profile consists of lithomarge clay. The X.R.D. analysis has revealed that the composition of the lithomarge clay is kaolinite. The initial pore solutions which were high in alkalies were gradually removed and the pH of the medium thereby lowered. Further up in the profile the downward percolating waters which have a high dissolved CO_2 content, increases the acidity of the medium . The removal of silica occurs at higher pH conditions. The removal of silica and montmorillonite increases the porosity and permeability of the medium. The montmorillonite transforms into kaolinite by the intra-crystalline leaching of the inter layer cations and tetrahedral silica layers.

The iron is mobile and is being transported upwards according to the mechanism postulated for the process of lateritization. Due to the slightly acidic medium in this band alumina becomes mobile and the mobility of silica at the same time starts decreasing.

The zone overlying the lithomarge is of laterite, where there is an enrichment of iron. Iron is stable under strongly oxidising conditions and therefore immoblie (Craig & Loughnan,1964). In the oxidising condition and the rainwater containing dissolved CO_2, the chemical balance of the pore solutions is affected. Ferrous iron is changed to the immobile ferric form and deposited in the form of laterite. Some silica is deposited while some is washed away. The alumina becomes mobile in the acidic medium and is washed away with the lowering of the water table. In short there is absolute iron enrichment, as a result of which ferallite and siallite components separate to form hematite and goethite at one end (represented by laterite) and kaolinite at the other (represented by band of lithomarge) (Craig and Loughnan, 1964).