Plate Tectonics

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Week - 02
Lecture - 07
Oceanic Crust- I

Okay friends, good morning and welcome to this class plate tectonics. So, up to now we are pretty sure that the crustal thickness and the crustal composition, the crustal physical and chemical properties are very much important. Because when we are talking about this plate interaction, we will talk about the lithosphere which mainly composed of the crust as a whole and the upper part of the upper mantle. That's why we are emphasizing on this crustal properties and crustal behavior with respect to temperature, pressure, gravity, composition, thickness and so on. So, today we will confine ourselves in the oceanic crust. Because most of this plate tectonic, the oceanic crust is destroyed and the oceanic crust is created and this continental crust which was formed in the Archean time, it remains as it is with some modification due to crustal evolution and deformation.

However, the oceanic crust it is newly created at the mid-oceanic ridge and it is recycled at the subduction zone and finally, it is forming a conveyor belt type system. That's why the oceanic crust understanding about its behavior, about its nature and its response to tectonic is very much important. So, the oceanic crust it is typically 6 to 7 kilometer thick and it lies the average ocean water depth about 4.5 kilometer.

So, this 6 to 7 kilometer thick is the average thickness I was talking, but it varies from ocean to ocean from tectonic setting to tectonic setting. For example, if you talk about this oceanic lithospheric system in the Pacific Ocean, this crustal thickness is somehow different than the Atlantic Ocean. So, on an average it is 6 to 7 kilometer thick and the thicker oceanic crust occur where the magma supply rate is anonymously high due to higher than the normal upper mantle temperature. We know this basaltic magma which is the chief component of the oceanic crust which is generated at the mid-oceanic ridge. and the mid-oceanic ridge, it is where this rate of spreading is less compared to the rate of magma eruption.

So, there will be accumulation of magma and there will be anomalously thick oceanic crust will be created. And similarly those area which are characterized by anomalously high upper mantle temperature that area represent this magma eruption and magma creation at the upper mantle system. Similarly there is thinner than normal crust it forms where this upper mantle temperature is anomalously low typically because of very low rate of formation. So, here there is an exception maybe exceptionally thick and there is an exception maybe exceptionally thin, but on an average this oceanic crust thickness it varies from 6 to 7 kilometer and very important thing is that this oceanic crust it is in isostatically balanced with the continental crust in a crustal setting. So, if you see this block diagram here we have the oceanic crust and this is the ocean water and this is the oceanic crust thickness and this is the continental crust thickness.

Now imagine the continental crust it is thicker as compared to oceanic crust because if you remember we are talking about this continental crust behavior the thickness of the continental crust about 40 kilometer. However compared to that this oceanic crust thickness is about 6 to 7 kilometer and it is composed of mainly basalt and this is granitic to granodiorite and anorthosite like that. So, that means here it is chiefly composed of the silicate minerals mostly the silica and here it is mostly the mafic minerals. So, it is felsic composition it is mafic composition. So, that's why its density is different.

Here this continental crust are less dense as compared to the oceanic crust which is high density. So, two different density material they are placed side by side. So, they are putting pressure on the low-lying asthenosphere. So, the continental crust being thick it is putting more pressure compared to this oceanic crust. So, that means if you see here this continental crust it is putting more pressure.

So, that is the asthenosphere it is going down compared to this asthenospheric situation below this oceanic crust. So, being it is thick and it is thin they are putting certain pressure here and here. However, if you are moving further down and further down for example, if you see here if it is moving further down there is a level at which the pressure created by this continental crust and the pressure created by this oceanic crust remain same and this level it is called the level of compensation. So, that means below the level of compensation the pressure exerted by this column of rock and the pressure exerted by this column of rock both are same. So, below that the pressure is considered to be hydrostatic and above that the pressure is considered to be lithostatic.

So, if you remember earlier class we are talking about this pressure and density of this oceanic crust and the continental crust where we are talking about the pressure remains constant after the level of compensation. So, this is the level of compensation below that the pressure increases hydrostatically. So, there would be at less dense crust at the mountain and high dense crust in the ocean basin. If you see here also it is 2.7 gram per centimeter cube however it is 3 gram per centimeter cube so, after the level of compensation the pressure remains constant or it increases at certain rate.

So, that's why this continental crust and the oceanic crust they are lying side by side by isostatically balanced manner and the oceanic crust it is divided into three principal layers. So, what are the three principal layers? The layer 1 it is mostly of sediment here you can see this is layer 1 it is sediment. So, here sediment it is the seabed surface material comprises unconsolidated deposit including terrigenous sediment carried out into deep sea by turbidity current. Along with that there will be pelagic deposit such as zeolite clays, calcareous and siliceous ooze and manganese nodule. That means the layer 1 which is composed of sediment it composed of both the terrigenous sediment which is carried out to this deep ocean basin by the turbidity current from the continental margin to deep ocean basin.

And other sediment which is the basin's own sediment it is precipitated calcium carbonate, precipitated silica, precipitated manganese nodule, precipitated iron nodule those are basin's own sediment deposited by chemical precipitation. Apart from that there is some sediment which has biogenic origin that is the ooze they are also contributing to that. So, it is a layer which is composed of sediment of different origin. Additional to that there are certain ocean basins which receives sediment from volcanic origin also there is volcano, volcanic ash it is falling down. So, volcano clastic sediments are also there.

So, by and large you can say sediments or the layer 1 on the oceanic crust it is derived from multisource. And here you can see these are some of these photographs that is the zeolite then the manganese nodule, then this is the clays and this is the carbonate sediment above the CCD and this is the pelagic sediment derived from the organics. So, all these sediments they are contributing as layer 1 into the ocean basin. Now the layer 1 is about 0.4 kilometer thick that is 400 meters thick.

However, it varies from place to place. Near to the continent if you see here near to the continent this layer thickness is more. However, if you are going to this mid oceanic ridge system either it is very thin or it is not present. So, though it is about 0.4 kilometer thick it is the average thickness and its thickness increases towards the continent and decreases towards the mid-oceanic ridge.

Similarly, the input with either this sediment with which is near to this continent it will be of clastic sediment or it will be non-clastic sediment and that depends upon the depositional environment and the Eh-pH of this water. Anyway the sedimentary thickness varies from continental side to this mid-oceanic ridge side and it progressively thickens away from the ocean ridge where it is thin or absent. So, it may or may not present here, but if it is present it is very negligible thing. And there is a systematic variation in the sedimentary thickness from different oceans and within that ocean at different parts also. For example, it is taken here the Atlantic, Indian Ocean and the Pacifics.

So, if you see here this sedimentary thickness is represented in this relief model. Here you see this area close to India that is the eastern margin of Indian subcontinent it is representing high sediment thickness. Similarly, those areas which are here they are representing the higher sedimentary thickness and mostly if you notice mostly the high sedimentary thickness it is around to the continent about to the continent. And particularly here you see we have Sundarbans delta, we have Krishna-Godavari delta, we have Mahanadhi delta all these major rivers are depositing their sediment to the eastern margin. That's why we have more sedimentary input compared to the western margin of Indian subcontinent.

Now compared to this Pacific, Atlantic and Indian Ocean the sediment thickness vary systematically. For example, this Pacific Ocean is the you can say this older one as compared to this Atlantic Ocean. So, it is obvious if it is older then means the sedimentation for millions of years. So, it should contain thicker sediment compared to the Atlantic. However, the reverse is true.

The oceans or the pacific ocean it is surrounded all sides by this trenches by the subduction zone. So, trenches that means like this, this is the subducting system and this is the abducting system and here is a trench. So, if this continent is here the sediment is being supplied by or transported by the rivers or anything. So, the sediments are mostly occupied at the trenches rather reaching to this deep ocean basin. So, that's why this periphery of this Pacific Ocean at it is surrounded by deep trenches the sediments which are derived from the continents they are this trench is behaving as a trap to accumulate the sediment here.

And that's why less sediment they are reaching deep inside this ocean. So, that's why the sedimentary thickness in the Pacific is less. However, if you compare to the Atlantic here, Atlantic all side it is bounded by the continent and here if you see all side it is bounded by the continent and there are major rivers they are depositing their sediments here. That's why the sedimentary thickness in the continental side or near to the continental side or overall in the Atlantic Ocean is more as compared to the Pacific Ocean. Now, layer 1 is sediment we all are convinced.

Now coming to the layer 2 which is mostly of volcanic origin and it is a rugged surface and mostly it is composed of pillow basalt. Pillow basalt, pillow basalt you know when there is basaltic eruption in subaqueous environment either it is in ocean basin, in the lake, river or it may be in the permafrost region, the ice sheet. So, there will be sudden cooling there will be quenching of this upper surface. So, it creates the pillows these are the pillows of a basalt. So, here the layer 1 is sediment, the layer 2 mostly it is composed of pillow basalt.

So, that means, the basaltic magma it is originating from here and once it is erupted here. So, due to this water content of the sediment and this ocean water it is suddenly quenching and forming some pillows. So, layer 2 is pillows, layer 1 is sediment. So, layer 2 it is of igneous origin basalt is olivine theolitic containing calcic-plagioclase and are very poor in potassium, sodium and other incompatible element. So, we know this basalt it is of basalt theo-olivine thiolite.

So, this Tholeiite it is a kind of basalt there is a compositional variation from this normal basaltic region and it is rich in iron and poor in aluminum. So, this tholeiitic basalt it is the typical characteristics of the mid-oceanic ridge basalt and that is poor in potassium, sodium and other incompatible element. Very little lateral variation in composition except at the volcanic islands. If you see here this volcanic islands are here. So, these are these hotspot magma derived or formed by the hotspot magma and this is the ocean ridge basalt.

Now you see the ocean ridge basalt it is behaving as a carpet it is lying on the asthenosphere the basaltic lava carpet it is on the asthenosphere which is representing the ocean basin. And the magma is derived from the mantle or this the upper part of this mantle. However, if you compare to the hotspot magmatism, the hotspot magmatism it is coming from D double prime layer it is around this mantle-core boundary. So, now imagine we have a compositional stratigraphy at one place a magma is derived from this level having certain composition and another place or the hotspot magma which is derived from this level it is compositional difference obviously. So, that's why except this is the hotspot magma or the volcanic islands all other part of this oceanic crust it is of uniform composition that is the olivine tholeiite and this layer have a variable thickness that is from 1 to 2.

5 kilometer and the seismic velocity varies from 3.4 to 6.2 kilometer per second that we have already discussed here. You see these are the hotspots and the hotspot magma is derived from D double prime layer. However, normal oceanic crust to form normal oceanic crust this asthenospheric magma it is deriving here at the mid-oceanic ridge.

So, that means two different levels the magmas are deriving. So, that's why it will of different composition compared to this that's why there is a variation at the oceanic islands. Now this layer 2 it is again divided into three sub layers. What are those sub layers? There is layer 2A it present only near the oceanic ridge and where the hydrothermal circulation of sea water occurs. Its thickness is about 0 to 1 kilometer and it is porous that is 30 to 50 percent porosity and it is rubbly nature and hence allow water to circulate.

Now if you concentrate yourself here you see at the mid-oceanic ridge system we have different faults inward facing fault or it may be inward and outward facing faults. So, either of this case so, these faults allowing water to percolate apart from that we have this pillow basalt and the pillow basalt they are highly porous there are high porosity and permeability among the different pillows. So, that's why anyway it is allowing the water to percolate down. So, this area is porous and permeable so, allowing this water to percolate down.

Now this represents layer 2A. Now away from that it is layer 2B which is formed by the filling of the porosity of the 2a by secondary mineral such as calcite, quartz and zeolite. Now if you are going away from this mid-oceanic ridge gradually we are going away from this hot environment hot magmatic environment. So, that means we are allowing the secondary minerals to precipitate. We have already porosity here we have faults, we have fractures and we have pore spaces within that pillows. So, once we are moving away from the system we are allowing this system to precipitate the secondary minerals.

Now the third layer that is 2C it is dominated by intrusives and grades into layer 3. So, that means layer 2A is porous, layer 2B the porosity is filled by secondary minerals, layer 2C it is intruded by the dikes or intrusive rocks. So, now if you see here this is layer 2A mostly the pillows you can see the pillows are lying here and within that pillows there are pore spaces through which water is percolating down. Now coming here this is 2B and the 2B if you see here this white colours they are the secondary minerals they are filled here the porosity is filled you see this is the pillow and around this pillow these are white colour minerals they are surrounding. Similarly this is a pillow and in between this pore space is filled by white colour minerals that is the zeolites or quartz and calcite like that.

Now this is these two photographs this is representing 2C that is the intrusives. So, you can see this within that pillows there are some intrusions they are the dike intrusions here within this pillow there are dike intrusions. So, there is clear cut distinction between this layer 2A, 2B and 2C. Then we are grading to layer 3 of the ocean basin or oceanic crust. So, layer 3 it is nothing it is the sheeted dikes it is the sheeted dikes.

So, if you see here these are this mantle material and from which this fractures they are surrounding fractures they are supporting this magmatic eruption. So, these are the conduits through which the magma is erupting and forming the pillow basalt at this level. So, the sheeted dikes are nothing these are the conduits these are the fractures which are allowing this magma to lift from this mantle to the surface or near-surface. So, these are solidified volcanic system vents these are the dikes they are the fractures which are filled with magma and it is solidified.

So, these are the dikes. So, this is representing layer 3 of this oceanic system. Here some field photographs you can see this is the sheeted dikes they are very closely spaced dykes and they are intruding into the basaltic pillows at the upper part and the lower part if you are moving down. So, this becomes coalescence and forming a sheet-type structure that means, very closely spaced dykes are there. Then we are landing to oceanic layer 4 that is the gabbroic layer. So, layer 1 is sediment, layer 2 is pillows, layer 3 is sheeted dyke, then layer 4 is the gabbro.

It is the chief layer of the oceanic crust and it is characterized by its plutonic foundation. Now it is the plutonic rock. So, gabbro that is a plutonic rock and gabbroic composition with few places of a serpentinized ultramafic mineral. Now here serpentinized. So, why serpentinized? Because we are allowing water to percolate down.

So, this due to percolation of water this ultramafic minerals they alter due to presence of water. So, finally, olivine and pyroxene which are the chief component of this basalt it is altered to serpentinite or serpentine. If you see here this black parts they are fresh basalt. However, if you see some part becomes grey and whitish they are nothing they are the altered product of this basalt they are the serpentinite or serpentine. So, that means, part of this basalt it becomes serpentinized due to alteration and this alteration in the presence of water and this water comes from this fractures through this pore spaces from the pillow basalt.

And this layer 4 or the gabbroic layer which is divided into 2 layers again. So, this is called layered gabbro the lower part it is called layered gabbro and the upper part it is called isotropic gabbro. So, layered gabbro and isotropic gabbro it names it says layered gabbro that means, it is a layer like material and isotropic gabbro that means, it is isolated patches are there. So, if you see this field photographs of this 2 layered gabbro here you can identify and distinguish different layers and isotropic gabbro that means, there is no layering occurs here. So, that's why these minerals particularly the mafics they are occurring within that dominated felsic as isolated patches.

So, now the question arises why this type of arrangement is there? Imagine there is a magma chamber inside and we are allowing the magma to cool down. So, the first mineral to crystallize the olivine the chrome spinels like this. So, if you remember this Bowen's reaction series here we are creating olivine and then pyroxene and amphibole like this. Similarly this side we are creating it is arorthite and here it is albite. So, that means, I want to say when we have this minerals which crystallize density is more.

So, they try to settle down to the bottom of this magma chamber. So, once they settle down so, they are forming layers. So, layer of rock is forming, layer of minerals which is mostly composed of ferromagnesium system ferromagnesium minerals like olivine, pyroxene they are settling down. Similarly this is the plagioclase is forming at the similar temperature.

at the similar temperature so, it is also settled down. So, they are forming layers. So, now imagine once this from this magma these components are removed due to crystallization. So, the remaining magma will be more felsic in composition. So, that's why those minerals which are forming that time.

So, they will try to settle down. However, that time this magma viscosity would have increased because the mafic part we have removed and the temperature we have reduced and this magma becomes more felsic that's why it will become more viscous compared to the initial composition of this magma. So, that's why these mafics which are crystallizing then though they try to settle down however due to viscosity of this medium high viscosity of this medium they will not able to settle down. So, they will be remain here and there where it crystallizes. So, that means we are creating a rock which is mostly of felsic but here we are having the mafics which are isolated occurring at isolated patches.

That's why it is called isotropic gabbro. So, the lower part it is the layered gabbro and the upper part it becomes the isotropic gabbro. Here some of the field photographs are also given this is the layered gabbro and these are the isotropic gabbros are there. The oceanic lithosphere forms at the mid-oceanic ridge and the widely accepted model proposed by Cann it describes hot asthenospheric material ascent buoyantly sufficiently rapid up to narrow zone to pass through the basalt melting curve and provides an interstitial melting of basaltic composition. Mostly if you imagine we have a asthenosphere and we are just separating the two plates away from each other and we are creating a narrow zone narrow rift environment and through this rift environment the magma is moving up.

This type of system it is called decompressional melting. So, decompressional melting is also it is called adiabatic melting. So, here we are removing two blocks away from each other. So, that we are decreasing the pressure. Once we are decreasing the pressure this it reaches the asthenospheric material it reaches to the melting point. So, that's why it starts melting and creating the magma here and the magma ascends through this fracture it is forming the sheeted dyke here and from this magma chamber we have this gabbros, layered gabbros and isotropic gabbros then we have this that is the sheeted dyke then this magma is erupting here creating this pillows.

So, this is a totally cross-section that can be imagined from here and the molten fraction increases in volume as the asthenosphere rises and eventually departs the parent material to ascend independently and produce the magma chamber within the lower part of the oceanic crust it is layer 4. So, here we are creating a magma chamber which is separated from this rest of this asthenospheric system because here in local we have reduced the pressure. So, we have decompressional melting we are creating in this region. So, we are creating a magma chamber here and here this is the schematic diagram how the system is doing. Here we have a magma chamber and this from this magma the mafic part first crystallizing and it is settling down crystal settle to the bottom of this magma chamber through non-viscous magma because at the initial time when the temperature was high and the magmatic composition was that much no differentiation or very little differentiation was there.

So, this magma is of mafic composition and it is of less viscous. However, once more and more magma more and more mineral is settled down differentiated the magma becomes viscous. So, the first mineral to crystallize in the magma chamber is the olivine and chrome spinel fall through this magma and form the basal layer of this dunite and peridotite occasionally chrome spinel. So, this is the initial stage where we are creating the layered gabbros. With further cooling pyroxene crystallizes and cumulates of a peridotatic layer. So, they are produced giving way upward to pyroxenite as the crystallization of pyroxene begins to dominate.

So, gradually the olivine the dunite is removed then the rest number in the Bowen's reaction series is olivine and pyroxene. So, after the removal of this olivine or the crystallization of the olivine the pyroxene starts crystallizing because the temperature gradually drops. So, we are getting a layer of pyroxenite there. So, ultimately plagioclase this side is also crystallized and layered olivine gabbro is formed. So, much of this residual liquid still volumetrically quite large then solidifies over a very small temperature range in the form the upper isotropic gabbro.

So, that means we have this isotropic gabbro here then we have layered gabbros here. A small volatile rich residual is this differentiation process consisting eventually of a plagioclase and quartz is the last fraction to crystallize. Sometimes it is intruded at the base of this sheeted dyke complex and that is called plagiogranite. So, that means now imagine we have a magmatic system from which we first created the layered gabbro then we created the isotropic gabbro then we have this peridotite, dunite, pyroxenite all these things this mafic components gradually we removed and the remaining magma become rich in volatiles rich in felsic components. So, after all this crystallization so that felsic part it creates the plagiogranite why it is plagiogranite because it is rich in plagioclase.

So, that's why the name plagiogranite is here. So, this plagiogranite it intrudes into the sheeted dyke complex if you see here this is the sheeted dyke complex they are the dyke and these are the plagiogranite part which is intruding at a tongue and apophysis or it is a large scale body you can find here at the base of this sheeted dyke complex. So, this is the complete topography or complete compositional system or compositional variation and lithological variation of a oceanic crust starting from the sediment then we have pillow basalt, then we have sheeted dikes, then we have isotropic gabbro, then we have layered gabbro and all through later this plagiogranite is cross-cutting. So, the plagiogranite being it is formed at the later part, but it is have intrusive relationship with some of this earlier formed litho units. So, plagiogranite are nothing they are tonalite composed of quartz and sodic plasioclase with minor mafic silicate. So, these are nothing with the tonalite why it is called tonalite this is light tone that's why it is called tonalite.

So, light tone is due to the felsic components and it is mostly composed of sodic plagioclase that is albite and with minor mafic silicates. They typically have granophyric intergrowths and maybe intrusive into layered gabbros that we have seen here some of this plagiogranite which is here some field photographs. This is the layered structure of this oceanic lithosphere starting from layer 1 which is sediment, then it is pillow basalt then we have this sheeted dyke then we have isotropic gabbro, then we have this layered gabbro, then it is cross-cutting by the plagiogranite. Thank you very much.