PLATE TECTONICS THROUGH TIME

The Lithosphere

The Lithosphere as a Thermal Boundary Layer — As discussed in the last lecture, the lithosphere can be thought of as a mechanically and chemically differentiated thermal boundary layer developed in connection with thermal convection in the mantle. The lithosphere is distinguished from the underlying “fluid” mantle — and from the cold thermal boundary layer in the last lecture’s idealized convection cell — not only by its lower temperature but also by its long-lasting strength (i.e. its ability to store up the energy released in earthquakes by deforming reversibly). In contrast to the cold thermal boundary layer in a fluid, the lithosphere acts as a stress guide: the force acting on a plate at any one place is transmitted throughout the plate, even over thousands of kilometers.

The thermal lithosphere, which is what geologists ordinarily mean by “lithosphere” in the generic sense, is comprised of the rock that is cool enough to behave in this way. Its lower limit corresponds to an isotherm (surface of equal temperature), commonly taken to be the 1200°C isotherm, which comes up almost to the sea floor at a midocean ridge, but may lie nearly 200 km beneath the middle of a continent. Archean terranes in the nuclei of continents commonly have a “keel” of anomalously cold mantle that extends even deeper.

The elastic lithosphere, in contrast, is comprised of rock that is cool enough to be not only rigid but elastic (i.e. to spring back to its original shape after deformation). The temperature cut-off for elasticity is around 500-600°C, depending on such things as the rock’s water content and structural integrity. Typically, the elastic lithosphere corresponds to the top one-third to one-half of the thermal lithosphere.

Lithosphere is a chemically differentiated thermal boundary layer in the sense that it is divided into the crust and the lithospheric mantle. Oceanic lithosphere, with its relatively thin (c. 6 km) basaltic crust, is less strongly differentiated in this way than continental lithosphere, with its relatively thick (30-40 km) crust, which is granitic on top and mafic, like oceanic crust, on the bottom. Cooler and thus more dense than the underlying rock, lithospheric mantle is gravitationally unstable. The crust, on the other hand, is light enough to be gravitationally stable. Nevertheless, oceanic crust is thin enough that it gets subducted along with the rest of oceanic lithosphere. As the depth, temperature and pressure increase, oceanic crust’s positively buoyant basalt and gabbro transform to dense, negatively buoyant eclogite. Granitic continental crust, on the other hand, is too thick and too light to be subducted bodily into the mantle. In some rifts and collisional mountain belts, continental lithospheric mantle and parts of the lower continental crust evidently delaminate from the upper crust and subduct.

The Lithosphere as an Evolving Mosaic of Plates — The lithosphere is comprised of rigid plates that average around 100 km in thickness. The worldwide mosaic of plates — today, about a dozen major plates and numerous smaller ones — evolves in consequence of continual generation and destruction of plate material, which is concentrated at plate boundaries. As discussed in Lab 1, these boundaries are of three types: rifts and midocean ridges, where plates diverge; subduction zones, where plates converge; and transform faults, where plates slide laterally past one another. Oceanic lithosphere accretes at midocean ridges and, to some small extent, in marginal basins behind island arcs; it moves laterally away from ridges and eventually is consumed at subduction zones, where continental lithosphere is formed as part of the subduction process.

The lithosphere tends to evolve progammatically: spreading ridges tend to keep spreading until they are subducted, and subduction zones tend to keep subducting until continents collide. One expression of this programmatic tendency is the Wilson cycle, the characteristic life cycle of an ocean basin like the Atlantic in which a new increment of continental crust is generated as the ocean opens and then closes.

1 Named for J. Tuzo Wilson, the cycle’s discoverer and one of plate tectonics’ pioneers.
The Wilson cycle can be thought of as beginning with the rifting apart of a continent. The rift develops into an ocean basin, and Atlantic-type continental margins develop on either side. In time, subduction zones develop along one or both sides of the ocean, in some places as island arcs, in others as Andean-type mountain belts. The two continents approach one another as the intervening ocean floor is subducted and eventually collide and become sutured together along a Himalayan-type mountain belt. In time, the cycle repeats itself, often with rifts beginning along the zone of weakness left in the lithosphere by the continental collision. Commonly, closing of the ocean basin includes one or more cycles of opening and closing of marginal basins behind island arcs.

As we will see in Lab 7 when we examine part of the original evidence for the Wilson cycle, present-day continental margins bordering the North Atlantic have been through about two cycles in the last 1 Ga, not counting secondary Wilson cycles involving island arcs and marginal basins.

Why the cyclicity? Why do new oceans to grow by the rifting apart of collisional mountain belts where earlier oceans closed? Evidently, extinct plate boundaries persist as zones of mechanical weakness in continental lithosphere for several hundred million years after they cease to be rifts or collision zones, and thus impart to the lithosphere a long-lasting “memory” by which past plate-tectonic arrangements for cooling the Earth’s interior influence future arrangements.

The Mechanism of Plate Tectonics

Cooling of Oceanic Lithosphere — As described in Lecture 9, heat flow from the mantle is the power behind plate motion. By far the largest amount of heat flow takes place through the ocean floor. Cooling and thermal contraction cause oceanic lithosphere to grow more dense than underlying mantle, and thus to grow gravitationally unstable. The apparent reason why fresh oceanic lithosphere does not subduct almost immediately is that the plate onto which it is accreted is like the hull of a boat: it will only sink if less dense underlying material can find a pathway onto the plate’s upper surface. Subduction zones are, in a sense, the holes in the hull.

The cooling process can be realistically modeled as the one-dimensional conductive cooling of an indefinitely thick body of rock whose upper surface is instantaneously chilled to the sea-floor temperature and maintained at that temperature ever after. As we saw in Lecture 9, the total amount of heat lost through a cold thermal boundary layer (i.e. oceanic lithosphere) depends on the time since cooling began (i.e. the oceanic lithosphere’s age) and the on the temperature difference across the lithosphere (i.e. the difference between the asthenosphere’s 1300° C temperature and the sea floor’s 4° C):

\[
\text{Cumulative Heat Loss } \propto \text{Temperature Difference } \cdot \text{(Lithosphere’s Age)}^{1/2}
\]

\[
\text{Heat Flow } = \text{Rate of Heat Loss } \propto \frac{\text{Temperature Difference}}{\text{(Lithosphere’s Age)}^{1/2}}
\]

2 Note that, in this connection, the oceanic lithosphere’s age means the time elapsed since its formation at a spreading ridge, not its age in years before the present. Thus, we might speak of Cretaceous lithosphere that is now 80 Ma old and of Jurassic lithosphere that was 80 Ma old at that time and is now 160 Ma old.

3 One sees why thermal conduction is ineffective relative to convection over the long term: the rate of conductive heat transport approaches zero with time.
The thermal lithosphere’s thickness (i.e. the depth to the 1200° C isotherm) depends only on its age:\(^4\)

\[
\text{Oceanic Lithosphere’s Thickness} \propto (\text{Lithosphere’s Age})^{1/2} \tag{10.3}
\]

**The Driving Forces** — The forces driving tectonic plates are somewhat more complicated than those driving the idealized thermal boundary layer considered as a model for them in the last lecture. There are two basic forces:

1. **Trench pull**, the force corresponding to the weight of subducting cold thermal plume in an idealized convection cell. Trench pull has two basic components: one associated with convection and proportional in magnitude to the convection cell’s vertical extent, and another associated with the elevation of the olivine-spinel phase transition. The first component’s magnitude is uncertain owing to uncertainty concerning the dimensions of convection cells. The second component arises from conversion of olivine to a more dense structure at a shallower depth owing to the subducting lithosphere’s relatively low temperature. Both components tend to pull the lithosphere down into the mantle, and both of their magnitudes are proportional to the lithosphere’s thermal thickness times the lithosphere’s density contrast with the asthenosphere, and to no other time-varying factors. Thus we have

\[
\text{Trench Pull} \propto \text{Temperature Difference} \cdot \text{Subducting Lithosphere’s Thickness} \tag{10.4a}
\]
\[
\propto \text{Temperature Difference} \cdot (\text{Subducting Lithosphere’s Age})^{1/2} \tag{10.4b}
\]

2. **Ridge Push** — As discussed in Lecture 8, midocean ridges are ridges because newly formed sea floor subsides as it cools in consequence of the rock’s thermal contraction and correspondingly increased density. Ridge push amounts to the gravitational sliding of a plate off a midocean ridge; it corresponds to the force that drives a sled downhill. The oceanic plate tends to slide off the ridge in response to the ridge push force, a gravitational body force that arises in consequence of the plate’s subsidence.

The sea floor’s subsidence is proportional to the lithosphere’s thickness times the difference between the lithosphere’s mean density and the asthenosphere’s (which is proportional to the temperature difference across the lithosphere), which means that

\[
\text{Subsidence} \propto \text{Temperature Difference} \cdot (\text{Lithosphere’s Age})^{1/2} \tag{10.5}
\]

The magnitude of the ridge push force is proportional to the subsidence times the plate’s thickness, which, as we find from (10.3) and (10.4), means that the ridge push force acting on a column of lithosphere increases in proportion to the square of the lithosphere’s thickness:

\[
\text{Ridge Push} \propto \text{Temperature Difference} \cdot \text{Lithosphere’s Age} \tag{10.6a}
\]
\[
\propto \text{Temperature Difference} \cdot (\text{Lithosphere’s Thickness})^2 \tag{10.6b}
\]

\(^4\) For simplicity, we neglect the oceanic crust. Present-day oceanic crust is buoyant enough to resist subduction but thin enough (c. 6m km) not to resist very much. Archean oceanic crust, on the other hand, the product of more complete melting of the mantle beneath midocean ridges, may have been as much as three times thicker, and thus less susceptible to subduction. However, being correspondingly more mantle-like in composition, Archean oceanic crust should have been less buoyant. The opposed effects on subductability should tend to cancel one another.
Notice that both of the driving forces are generated by *oceanic lithosphere* (as opposed to continental lithosphere).

In theory trench pull should be at least an order of magnitude stronger than of ridge push — strong enough to actually pull a subducting plate apart. In practice, the subducting slab’s breakup in response to high stress levels and the sublithospheric mantle’s resistance to flow probably limit the trench pull actually delivered to the rest of the plate to the magnitude of ridge push.

**Plate Tectonics Through Time**

The lithospheric plates can be thought of as the cooling fins on the endogenic heat engine. The last lecture discussed the engine’s operation and evolution from the standpoint of individual convection cells. Here, we examine how the evolving mosaic of plates adjusts itself to the whole engine’s heat-disposal demands.

**Interrelationship of Mantle Cooling and the Plate Mosaic** — Equation (10.2) gives the approximate relationship of the total oceanic heat flow— the main power behind tectonics— to the difference between deep-mantle and sea-floor temperatures and the oceanic lithosphere’s mean age:

\[
\text{Total Heat Flow} \propto \frac{\text{World Ocean's Area} \cdot \text{Temperature Difference}}{(\text{Ocean Floor's Mean Age})^{3/2}}
\]

\[
\text{Ocean Floor's Mean Age} \propto \left[\frac{\text{World Ocean's Area} \cdot \text{Temperature Difference}}{\text{Total Heat Flow}}\right]^{2}
\]

The ocean floor’s mean age depends on the plates’ average lateral extent as well as their average rate of movement. Expressions (10.7) and (10.8) thus show a fundamental connection between radiogenic heat production in the mantle (which determines the long-term average heat flow) and the overall organization and dynamics of the worldwide mosaic of plates.

The difference between deep-mantle and sea-floor temperatures evidently has decreased relatively little — from as much as 1600° C to about 1300° C — since the Early Archean, as shown by komatiites with eruption temperatures as much as 300° C higher than basalts in corresponding tectonic settings today. Abundant indications of liquid water and life in sedimentary rocks of all ages show that sea-floor temperatures have remained more or less the same.

**Changes in the Mosaic’s Design** — Because oceanic lithosphere’s mean age at subduction varies inversely with the square of the heat flow, the plate mosaic must have changed substantially with the steady, more or less exponential decrease in radiogenic heat production to about half its Archean value (see Lecture 9). Over the long term, oceanic plates’ mean age must have increased as radiogenic heating fell off. For instance, if the heat flow were twice what it is today, as it must have been in the Early Proterozoic or thereabouts, oceanic lithosphere’s mean age should be one-quarter of its present 60 Ma value, all else being equal.

How might the Precambrian oceanic lithosphere have accomodated the higher heat flow? The two basic options are an increased spreading rate and a longer midocean ridge system’s length, as we find by rewriting (10.8) with the sea floor’s mean age expressed in terms of ocean basins’ mean width divided by the mean spreading rate, and with the world ocean’s area expressed in terms of ocean basins’ mean width times the midocean ridge system’s length:
Mean Spreading Rate · Mean Basin Width · (Ridge System’s Length)$^2$

$$\propto \left[ \frac{\text{Total Heat Flow}}{\text{Temperature Difference}} \right]^2 \quad (10.9)$$

Ocean basins’ mean width, but not the ridge system’s length, depends on the amount of continental crust. One sees that a 10% change in ridge length has proportionately more effect on the lithosphere’s heat disposal than a 10% change in mean spreading rate or mean ocean-basin width.

Was higher heat flow in the past accommodated by correspondingly faster sea-floor spreading, by a longer midocean ridge system, by some combination of the two, or simply by wider oceans in the absence of present-day amounts of continental crust? As the accompanying diagram shows, apparent polar wander paths for the Late Archean and Early Proterozoic (which is about as far back as reliable polar wander paths can be reconstructed) indicate that terranes were moving latitudinally at no more twice their present-day rates, or roughly half the rate expected from the higher heat flow at that time. Wider oceans (a consequence of lesser amounts of continental crust at the time) and a slightly longer midocean ridge system (amounting to perhaps one or two more major plates) seem to account for the rest of the difference in rates of plate movement.

**Changes in the Mechanism of Plate Tectonics** — Consider the forces on oceanic lithosphere that has just reached a subduction zone. Figuring that its age, on average, should be twice the mean age of oceanic lithosphere at the time, we find the relative magnitudes of the forces driving plate motion from (10.5)-(10.7)

$$\text{Ridge Push} \propto \frac{\text{Temperature Difference}}{\text{(Total Heat Flow)}^2} \quad (10.10)$$

$$\text{Trench Pull} \propto \frac{\text{Temperature Difference}}{\text{Total Heat Flow}} \quad (10.11)$$

One notes a seeming paradox: The more powerfully plates are driven by heat flow, the less forcefully they are driven by their own gravitational instability.

Plate tectonics may indeed have worked somewhat differently in the past, when heat flow was higher, the ridge push and trench pull driving forces were weaker, and ridge push was relatively less important. As we have already seen in the last lecture and in Labs 1 and 2, the 200-300$^\circ$ C decrease in the mantle’s overall temperature over the 4 Ga course of geologically recorded history has made a definite but none too profound change in the character of igneous rocks generated at plate boundaries. The next two lectures explore these consequences further as they examine the historical evidence in greater detail.

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5 The time-trend in the amount of continental crust has been controversial for decades. As the next two lectures will discuss, the isotopic evidence on the cycling of material through crust and mantle isotopic reservoirs appears to indicate that significant amounts of continental crust were generated only after the mid-Archean, and that the standing amount of continental crust has been more or less constant through the Phanerozoic.
Study Questions

0. Continuing Questions: Why are there so many tectonic plates? Have there always been about a dozen major ones? Have they always moved as fast as they do today? (Lectures 1, 2, 9) How can we tell? (Lecture 7)

1. What is the thermal lithosphere? …the elastic lithosphere? What are oceanic lithosphere and continental lithosphere, and how do they differ? What distinguishes the crust and mantle? …the lithosphere and asthenosphere? What is the significance of describing the Earth’s layering in these two different ways?

2. What is the Wilson cycle? What does it mean to say that continental lithosphere preserves a “memory” of tectonic plates’ past arrangements that can influence future arrangements?

3. What forces drive the tectonic plates? What is ridge push? …trench pull? What factors determine the magnitude of the driving forces? How have the determining factors changed through time, how have the forces changed in magnitude, and to what effect?

4. What does it mean to say that the lithosphere’s tectonic plates are tailored to dispose of the mantle’s heat? What does it mean to say that rates of plate movement are balanced against the length of the midocean ridge system, the number of tectonic plates, and the extent of continental crust? How has continental growth and radiogenic heat production influenced the balance, and how has the balance changed? What do measured rates of apparent polar wander indicate about the arrangement of continents, oceans, midocean ridges, and tectonic plates on the Archean globe?