GEOL311: Review sheet for Final exam

The final exam will cover ALL topics (1 through 15).
Please bring a compass – there WILL be a Mohr circle question

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Topic 11 – Fabrics: foliations and lineations

11.1 Introduction
A fabric is a geometrical arrangement of component features in a rock, viewed on a scale large enough to include many examples of each feature.
Primary fabrics form during the formation of the rock, e.g. bedding
Tectonic fabrics form as a result of tectonic deformation

11.2 Terminology
A random fabric shows no preferred orientation (e.g. undeformed granite)
A non-random fabrics can be continuous (spaced at less than ~1mm) or discontinuous (spaced at more than ~1mm, visibly discontinuous)
Rocks with a penetrative tectonic fabric are called tectonites. The two main types of fabric are planar (foliations) and linear (lineations):
S-tectonites have a dominantly planar fabric (S = surface)
L-tectonites have a dominantly linear fabric
LS-tectonites have both planar and linear fabrics, with neither dominant
May have several generations of fabrics

11.3 Foliations
Progression with increasing temperature from cleavage to schistosity.
Cleavage is a secondary fabric element (forms subsequent to formation of the rock), formed at low temperatures, that imparts a tendency for the rock to split along planes (but note that cleavage forms without loss of cohesion).
(i) disjunctive cleavage is defined by an array of subparallel cleavage domains (where original fabric and composition have been changed by pressure solution), separated by microlithons (relatively unchanged) Also known as pressure solution cleavage or stylolitic cleavage. Forms by removal of calcite or quartz grains leading to a concentration of clay minerals.
(ii) pencil cleavage is a result of a tectonic fabric (cleavage) superimposed on a primary fabric(bedding) in clay-rich rocks, which allows the rock to break into pencils.
Pencil cleavage may form as a prelude to slaty cleavage:
(iii) slaty cleavage : clays throughout the rock have the same tectonically induced preferred orientation, and there are no microlithons visible to the naked eye. Strain ellipsoid is oblate and parallel to the fabric. Forms at lowest grade metamorphic conditions, with some new mineral growth.
(iv) phyllitic cleavage is a result of higher grade (greenschist) metamorphism. New mineral growth has a strong preferred orientation if it occurs in an anisotropic stress field. Phyllites look like slates only shinier.
(v) crenulation cleavage forms when a rock is compressed at a low angle to an existing slaty cleavage. May be symmetric or asymmetric. Crenulation cleavage is usually S2, after the slaty cleavage (S1) has overprinted bedding (S0).
Pressure solution may occur simultaneously, in which case a mineralogical differentiation occurs, with quartz concentrated in hinges and micas on the limbs of kink folds. If complete, results in transposition of the old fabric into a new orientation. (vi) schistosity is a result of medium grade metamorphism, where new minerals such as garnet can grow. They may grow up to several cm, and muscovite, chlorite and biotite are also coarser. There are no clay minerals left. (vii) gneiss is a metamorphic rock in which foliation is defined by compositional banding. This may occur by (a) inheritance, e.g. from layered sediment, (b) fabric transposition (e.g. stretched isoclinal fold limbs), (c) metamorphic differentiation, where diffusion processes result in alternating mafic (Fe, Mg-rich) and felsic (Si, Al, and alkali-rich) bands, and (d) lit-par-lit intrusion, where melt is injected as thin sills along weak planes in the protolith.

11.4 Tectonic interpretation of foliations
In many folded regions there will be an axial planar cleavage, formed approximately normal to \( \sigma_1 \), and parallel to the axial planes of the folds. Sometimes there may be cleavage refraction, where cleavage in more rigid (more competent) layers retains its originally high angle to bedding. Cleavage refraction gives a tool for determining sense of shear.

11.5 Lineations
A lineation is a fabric element that can be represented by a line. (i) Form lineations are manifestations of the occurrence of other structures, e.g. (a) fold-hinge lineations and crenulation lineations (hinge lines of folds), (b) rods (detached fold hinges where limbs have thinned to zero) (c) mullions (cusp-like contacts between rocks of different competency), (d) boudins (sausage-shaped lenses of relatively rigid layers in soft matrix), (e) elongate objects e.g. stretched pebbles. (ii) Surface lineations are found on planar surfaces such as bedding planes: (a) intersection lineations occur at the intersection of two planar fabric elements (e.g. bedding-cleavage intersection), (b) slip lineations form on fault surfaces, or on bedding planes during flexural folding, and are parallel to the slip direction. Fibre lineations form when vein mineral fibres precipitate via crack-seal. (iii) Mineral lineations are defined by preferred orientation of mineral grains. May be due either to growth in a preferred orientation (controlled by differential stress field) or due to rotation of elongate grains towards a principal strain direction during progressive deformation.

11.6 Rock anisotropy
Foliations and lineations cause rocks to be strongly anisotropic. Detected by: speed of sound waves in different directions, magnetic anisotropy, etc.

**Topic 12 – Shear zones**

12.1 Introduction
A ductile shear zone is a tabular band of definable width in which there is considerably higher strain than in the surrounding rock.
Note - no single through-going fracture; boundaries are often gradational; displacement of one side relative to the other (simple shear).

Consider a deep fault that penetrates below the brittle regime of the crust:
Near surface brittle processes produce gouge and/or breccia
Deeper cataclasis (macroscopically ductile, microscopically brittle)
Below ~ 12 km quartz can now behave plastically ($T \geq 300^\circ C$)
ductile processes occur in quartz-rich rocks

The brittle-plastic transition usually occurs at about 10 to 15 km depth, but note:
(a) depends on temperature, and hence the geothermal gradient,
(b) depends on strain rate
(c) depends on mineralogy and composition

Shear zones are regions of distributed deformation and may be small (a few mm) or very large (kilometres wide)
They usually contain strong foliation, which is most intense in the zone of greatest shear. Mineral lineations form parallel to the slip direction (topic 11):

12.2 Mylonites
Mylonites are fine-grained rocks formed in shear zones, whose small grain size compared to the host rocks results from crystal-plastic processes, particularly dynamic recrystallization (remember topic 9!)
Marble mylonites and quartz mylonites form at lower temperatures than quartz feldspar: the brittle-plastic transition depends on the predominant mineral in the rock, as well as temperature and strain rate.

12.3 Shear sense
To find the sense of slip of a fault we have only to match up displaced beds or other markers. To do this in a shear zone, where such markers are usually absent, we use kinematic indicators, or shear sense indicators.
The optimum orientation for examining shear sense is found on a plane normal to foliation and parallel to lineation
There are several different types of kinematic indicators.
(i) grain-tail complexes form as material recrystallizes or precipitates in the flow field exerted by a porphyroclast (a large mineral which grew prior to deformation and behaves mostly as a rigid body) Sigma-type and delta-type.
(ii) disrupted grains form when more rigid grains fracture (note that the shear zone is still ductile if deformation occurs dominantly by crystal-plastic processes). Includes mica fish, synthetic and antithetically sheared clasts.
(iii) foliations may also give shear sense, where an earlier foliation ($S_n$) is sheared along a more recent foliation ($S_{n+1}$). Two types, S-C fabrics and extensional crenulation cleavage; both geometrically resemble miniature shear zones within shear zones.
(iv) transposed folds may give a misleading shear sense, particularly where there have been very high shear strains. Fold transposition can occur as asymmetrical folds grow and become refolded so that the limbs are parallel to layering. In extreme cases, fold limbs become stretched and attenuated to the point that they are either invisibly thin, or they form boudins with boudin necks separating limbs and hinges. In the outcrop the limbs will appear parallel to layering (and may be mistaken for part of the layering) – only finding the rootless isoclinal fold hinges will demonstrate that fold transposition has occurred.
Topic 13 – P-T-d-t paths

13.1 Introduction
Metamorphism is the mineralogical, chemical and structural adjustment of solid rocks to physical and chemical conditions imposed at depth. Changes in mineral assemblage (the constituent minerals in a rock) are the most obvious results of metamorphism. These changes occur in response to pressure (P) and temperature (T), over a period of time (t), so the metamorphic history of a rock is called its P-T-t path. In recent years, metamorphic petrologists have started to include a fourth variable, deformation (d), recorded in the growth of new fabrics (recall topic 11). This allows construction of P-T-d-t paths, and correlations between tectonic events (giving rise to structures), thermal events (recorded by changes in mineral assemblage), and time (recorded by geochronology).

13.2 Field observations
In the field we try to unravel sometimes complex histories, such as polyphase deformation, by first working out how many events occurred, and then their relative timing relations (using cross-cutting relationships).

13.3 Thin section work: porphyroblasts
By looking at the relations between porphyroblasts and tectonic fabrics, we can work out whether the mineral grew before, during or after deformation (pre-kinematic, syn-kinematic, or post-kinematic): Sometimes inclusions in porphyroblasts will contain minerals from a previous mineral assemblage, which allows us to tell whether temperature (and pressure) has increased or decreased between the earlier assemblage and the new assemblage which the porphyroblast is part of. Remember the difference between a porphyroclast and a porphyroblast! Also remember the names of rock types at increasing metamorphic grade e.g. sandstone becomes quartzite, limestone becomes marble, shale becomes slate then phyllite then schist then gneiss, etc.

13.4 Geothermobarometry
Geothermobarometry is the quantitative determination of past pressure-temperature conditions. Distinguish between prograde path (burial and heating) and retrograde path (exhumation and cooling). Anticlockwise P-T paths are found in collisional belts, where crustal thickening or overthrusting lead to a rapid increase in pressure. Temperature then increases more slowly due to thermal relaxation, as the geotherm relaxes towards its equilibrium value.

Topic 14 – Earth structure

14.1 Layers of the Earth
Defined by seismic velocities and bulk compositional layering. Know the composition of oceanic crust, continental crust, mantle and core. Know average thickness of oceanic crust (7 km) and continental crust (35 km)
The Moho is the jump in seismic velocities, approximately the crust-mantle boundary. Very distinct in oceans, less so under continents.

14.2 Rheology of the Earth
Lithosphere: exhibits flexural rigidity on geological timescales (has an "effective elastic thickness"). Will flow on long timescales, but at rates slower than geological strain rates. Heat transferred by conduction. Base defined by 1280 °C isotherm, i.e. thermal boundary (may vary in space and time).

Asthenosphere: Exhibits no flexural rigidity. Heat transfer by convection (crystal plastic deformation and diffusion) although it is still solid (transmits S waves). Low velocity zone in upper asthenosphere below oceanic regions, may be due to presence of partial melt.

14.3 Tectonic Plates
There are seven major plates and 12 or more minor plates. Plates can be entirely oceanic (e.g. Pacific) or contain both continental and oceanic lithosphere (e.g. North American). Oceanic plates consist of lithosphere only. Continental plates may also drag some asthenosphere with them, since there is no low velocity zone beneath continents. The combined thickness of material that moves together in continental regions is called the "tectosphere".

14.4 Plate boundaries
Plate boundaries are sharp and well defined between oceanic plates (e.g. mid-Atlantic ridge; Mariana trench) but may be diffuse in continental regions (e.g. Tibetan plateau). This is because oceanic crust is stronger than continental crust (olivine vs. quartz-feldspar). Divergent boundary = mid-ocean ridge, Convergent boundary = subduction zone or collision zone, Transform boundary = strike-slip fault, (e.g. San Andreas Fault)

14.5 Triple junctions
Combinations of ridge, trench, transform boundaries. Stable (basic geometry can exist for a long time although absolute position of triple junction migrates with time) or unstable. Triple junctions can evolve from one type to another.

14.6 What drives plate tectonics?
Mantle convection driven by Earth’s internal heat. Plate movements probably more influenced by “ridge push” at divergent boundaries and “slab pull” at convergent boundaries.

Topic 15 – Condensed Plate Tectonics

15.1 Extensional tectonics - Rifts
Rifts are regions of continental lithosphere undergoing extension. Successful rifting leads to sea floor spreading (and passive margin development). Unsuccessful rifting produces a basin, either above or below sea level (e.g. Basin and Range, North Sea). Cross-section of a rift: asymmetric, with tilted fault blocks either parallel (bookshelf model – space problem) or listric faults (no space problem, forms rollover anticline). Both types have been observed in rift systems.
Crustal scale models of rifting: pure shear (homogeneous lower crustal stretching), or simple shear (detachment cuts to base of lithosphere, explains asymmetry). Probably a combination of the two in most cases.

15.2 Extensional tectonics – Metamorphic core complexes
If extension of the hanging-wall during rifting is extreme, unloading and isostatic compensation may result in exposure of the footwall. This typically contains mylonitized footwall rocks overprinted by a chlorite breccia at the detachment, and overlain by isolated hanging-wall blocks which have undergone brittle deformation. This is called a metamorphic core complex, and is common in the Basin and Range province of the southwestern US.

15.3 Extensional tectonics – Rift-related rock assemblages
Igneous rock assemblage associated with rifting is a bimodal volcanic suite: Mafic rocks like basalts form from decompression melting of upwelling asthenosphere. These may be emplaced as dikes and sills, or underplated. If a mantle plume coincides with the rift, a flood basalt province will probably result (e.g. Parana, Columbia Plateau). Silicic rocks like rhyolites form from partial melting of the crust as basalts are injected.

Sedimentary rock assemblage associated with rifting:
(a) initially basin is above sea level, fills with non-marine clastics deposited in alluvial fans, (b) also may be lakes which may have high salinity, so gravels, sandstone, siltstone, shale and possibly evaporates, (c) Once below sea level, the shallow sea will usually deposit a thick evaporite sequence which is subsequently overlain by marine carbonates and clastics.

If rifting is unsuccessful, rift sediments act as a load that bends down the rift margins during thermal relaxation and subsidence, leading to formation of “steer’s head basin”:

15.4 Extensional tectonics – Passive Margins
If rifting is successful, the result will be a passive margin, e.g. East coast and Gulf coasts of North America, W coast of Europe, all sides of Antarctica...

Note that passive margins are NOT plate boundaries.

Post-rifting, hot thin lithosphere cools and sinks to form a passive margin basin. This fills with sediment, and loading causes the basin floor to sink further. Extensional faulting may occur in the sedimentary pile even once faulting has ceased in the underlying basement.

15.5 Extensional tectonics – Causes of rifting
1. Thermally activated rifts overly mantle plumes (large topographic expression, up to 2 km high and 1000 km diameter, combined with heating and weakening of lithosphere results in spreading)
2. Drift-activated rifts. These may initiate due to two sides of a continent being pulled apart by plate-driving forces (i.e. slab-pull since ridge-push won’t help much). Especially likely if continent contains a young hot weak orogen.
3. Gravitational collapse of weak, overthickened crust
4. Transtensional pull-apart basins on strike-slip faults
5. Back-arc extension at convergent margins
15.6 Contractional tectonics – convergent margins
Once oceanic lithosphere becomes negatively buoyant (after about 200 million years or less) it will sink into the asthenosphere and pull the slab down behind it, i.e. it is subducted. As it heats up, sediment and metasomatized oceanic crust will dehydrate. The overlying asthenospheric wedge is then fluxed with fluids and will partially melt to produce magma, which rises to the surface and forms an arc of volcanoes. If it is subducted under another oceanic plate this is an island arc (e.g. Tonga). If it is subducted under the edge of a continental plate this is a continental arc (e.g. Andes).
On a sketch you should be able to label the peripheral bulge, trench, accretionary wedge, volcanic arc, and back-arc region.
Buoyant lithosphere (including continental lithosphere, oceanic plateaux, island arcs and some spreading ridges) cannot be subducted. Instead, a collision zone develops and subduction ceases (the downgoing slab keeps going and usually breaks off; they have been imaged sinking deep into the mantle).

15.6 Contractional tectonics – continental collision
Always initiates with convergent tectonics. May get huge orogenic belt, (e.g. Himalaya) or just get suturing if driving forces stop when continents collide.
First, the initial interaction results in extension (and possibly sub-aerial exposure) of passive margin as it bends over peripheral bulge.
Next, subduction is aborted and collision begins. The passive margin is carried into the trench where it is unconformably overlain by arc turbidites. A fold-thrust belt initiates in the passive margin sediments and propagates toward the foreland, initiating a foreland basin ahead of it (flexure due to loading).
The passive margin is inverted and ophiolite may be emplaced at the site of the suture (plate boundary at the surface but usually transported a long way towards the foreland relative to the plate boundary in the mantle)
Crustal thickening and high-grade regional metamorphism in the internal zone of the orogen. Faulting in the external zone (fold-thrust belt).
Note that a fold-thrust belt may occur on both sides of the orogen (concept of a doubly-vergent orogen), but should be able to tell which side subducted because the volcanic arc will be on that side of the orogen.

15.7 Contractional tectonics – Fold-thrust belts
These are usually underlain by a basal detachment (floor thrust), and may be overlain by an upper detachment (roof thrust). Imbricate fan vs. duplex. Faults may be emergent (intersect surface) or blind.
Note thrusts frequently contain frontal, lateral and oblique ramps
New thrusts typically develop in a break-forward sequence (progressively younger towards the foreland), although out-of-sequence faults may develop, as can backthrusts.
Thrust-related folds: fault-propagation folds and fault-bend folds
Fold-thrust belt mechanics: remember
effective normal stress = normal stress – pore fluid pressure
Frictional resistance to sliding is directly proportional to effective normal stress
So high fluid pressure allows easy sliding on pre-existing fracture

15.8 Contractional tectonics – Orogenic collapse
Once the crust becomes very thick, it will heat up. It may even partially melt.
Hence it gets weaker, and more force is required to support the high mountains that were originally produced by isostasy. If the boundary force does not increase then gravity will drive spreading, at least of the upper crust. Another factor is that the entire lithosphere gets thickened, not just the crust. Since the lithosphere may be thickened by two times or more (say 200 km thick), now have cold lithosphere surrounded by hot asthenosphere. This arrangement is unstable and the lithosphere can delaminate (drop off). This leads to a very rapid isostatic rebound and uplift of the crust. Asthenosphere upwells to replace the lithosphere and partially melts, which in turn heats the crust through intrusion and could lead to further crustal melting.

15.9 Strike-slip tectonics - Introduction
Transfer faults: Displacement can be constant along the length of the fault, and total displacement can be much greater than the length of the fault. e.g. Transform faults (plate boundaries) offsetting segments of a mid-ocean ridge (fig. A). [Note that fracture zones that extend beyond the edge of the transform faults are not sites of active faulting, nor are they plate boundaries.] Can also occur on the mesoscopic scale, e.g. connecting two veins as they open. Transcurrent faults are NOT plate boundaries. They initiate and grow from a point, so they die out along their length. Maximum displacement is always shorter than their fault length. Often terminate in fault splays (horsetails). Distributed deformation: Continental crust is quite weak, so deformation is often distributed across a broad zone. Major strike-slip fault zones often contain many en echelon synthetic strike-slip faults at a low angle to the main fault, and sometimes antithetic strike-slip faults at a high angle to the main fault. Folds or thrust faults, and normal faults, also form at specific orientations with respect to the strain ellipse:

15.10 Strike-slip tectonics - Transpression and transtension
Can either be partitioned into discrete dip-slip and strike-slip zones, or will result in formation of flower structures: negative if transtensional, positive is transpressional. Note that transpression and transtension usually result from moving past restraining bend or releasing bends. As one side of the fault moves from one to the other it is likely to reactivate and invert a pre-existing flower structure, so history can be very complex.

15.11 Strike-slip tectonics - Tectonic settings of continental strike-slip faults
1. Oblique convergence and collision. May result in partitioned deformation, e.g. Wrangellia now found between Idaho and Alaska.
2. Escape tectonics (slip-line tectonics). e.g. extrusion of China
3. Lateral ramps in fold-thrust belts
4. Accommodating different amounts of rifting., e.g. Garlock Fault separates the Basin-and-Range from the Mojave Block.

15.12 Summary
The plate tectonic cycle repeats itself as one sort of plate boundary turn into another sort. Collisional orogens are often the sites of future rifting because the crust is overthickened, heats up and is weakened. Rifted margins are inevitably the future sites of convergence and subsequent collision.