Alpine Orogeny

The geologic development of the Mediterranean region is driven by the Alpine-Himalayan orogeny, a suturing of Gondwana-derived terranes with the Eurasian craton. In broad terms, this is a Mesozoic and Cenozoic convergent zone that extends from the Spain to Southeastern Asia and may extend along the southwest Pacific as far as New Zealand (Rosenbaum and Lister, 2002). The Alpine orogeny was caused by the convergence of the African and European plates (Frisch, 1979; Tricart, 1984; Haas et al., 1995) with peak collisional phases occurring at different times: Cretaceous in the Eastern Alps and Tertiary in the Western Alps (Schmid et al., 2004).

Note: prior to the opening of the Paleotethys sea, the Variscan orogenic belt developed in central Europe then the Laurussian and Gondwana converged in the Devonian and Late Carboniferous. Although the location of the suture is not clear, the orogenic belt was extensive, running from the Bohemian Massif to the Alpine-Carpathian-Dinarides belt.

The Paleotethys sea existed in the Triassic but closed in the early Mesozoic due to convergence along the Cimmerian (and Indosinian) suture zone. The Paleotethys (or Tethys I) has been described as a wedge-shaped ocean that opened to the east, separating Eurasia from Africa, India, and Australia (Laurasia and Gondwana). Very little evidence of the Paleotethys exists today which has caused some to question its existence (Meyerhoff and Eremenko, 1976).

The Tethys opened as Pangea broke up in the Early Jurassic and Africa moved east relative to Europe. There is abundant evidence for the Tethys basin including ophiolite sequences (including the Troodos Massif of Cyprus) and pelagic sediments. A Jurassic-Cretaceous spreading center was probably active in the eastern Mediterranean (Monod et al., 1974). Plate kinematics suggests 2,000 kms of sinistral movement. This sinistral transtension continued until the Late Cretaceous (80 Ma) when movement reversed and became dextral transpression. Plate kinematics models for the motions of Africa and Iberia relative to Europe indicate that convergence commenced between 120 and 83 Ma (Cretaceous Normal Superchron) and there were two periods of rapid convergence, during Late Cretaceous and Eocene-Oligocene. Relatively slow convergence occurred during the Paleocene and since the Early Miocene. The Alpine orogenic collision and the Early Miocene reduction may have led to extension in the Mediterranean back-arc basins. Paleocene slowing may have been due to Convergence can be separated into five phases (Rosenbaum et al., 2002):
1. Late Jurassic – Early Cretaceous left-lateral strike-slip.
2. Late Cretaceous convergence.
3. Paleocene quiescence.
4. Right-lateral strike slip motion.
5. Eocene – Oligocene convergence.

This closing of the Tethys sea led to the Alpine orogeny. Widespread flysch deposits record this closure that probably culminated in the Eocene. Only the eastern Mediterranean Ionian and Levantine basins remain as possible relics of the Mesozoic ocean basin.

A series of post-compressional basins developed superposed on the orogenic zones (Hsu and Bernoulli). The western Mediterranean and Aegean basins developed after the Alpine compressional event. The Balearic basin developed during the late Oligocene or early Miocene due to extension, the Tyrrenian basin is somewhat younger, and the Aegean basin has experienced considerable Pliocene-Quaternary subsidence (Hsu and Bernoulli).

More to come: Paleotethys, Variscan orogeny, Tethys, Late Miocene Messinian salinity phase.

*Plate kinematics models for the motions of Africa and Iberia relative to Europe (Rosenbaum et al., 2002).*
Geology of the Barcelona Region

The city of Barcelona is located on the Littoral Plain of the Catalan Coastal Range (Littoral Range), between the Basós and Llobregt rivers. The NE-SW trending topography that make up the Catalan Coastal Range is the result of an extensional fault system that has developed since the late Oligocene with most extension occurring in the Miocene (Santanach et al., 2011). The extensional regime remains today. The Catalan Coastal Ranges are the northern, onshore margin of the València Trough. This is an extensional basin that extends 250 kms along the eastern margin of the Iberian Peninsula. The Barcelona Basin is the portion of the València Trough just offshore of the city. A series of half-grabens step down from the Ebro Basin, north of Barcelona, to the València Trough. Tension that produced the Trough also caused uplift of the Catalan Coastal Range (Santanach et al., 2011).

Many of the extensional faults are reactivated compressional faults that formed in the Paleogene due to the Pyrenean orogeny compression. At that time, the thrust system uplifted the Catalan Coastal Range and part of the València Trough. These features may well have been earlier Mesozoic extensional faults that developed along the western edge of the Tethys. It seems to be a story about extension and compression that reactivates zones of weakness. The Paleogene uplift of the Catalan Coastal Range resulted in erosion of the Mesozoic and part of the Variscan basement (Catalan-Balearic massif) and deposition in the Ebro Basin (Santanach et al., 2011).

Uplift and erosion of the footwall of the main extensional fault continued during the Neogene, exposing pre-Variscan Paleozoic rocks and Carboniferous/Permian intrusives (Santanach et al., 2011). These units
are now exposed in the Littoral and Prelittoral ranges, including the hills of Barcelona. In the case of the Barcelona hills, they represent part of the hanging wall of the Collserola fault.

The old town of Barcelona is located on the Barcelona Plain, or Littoral Plain. This relatively flat region consists of Pleistocene alluvial fans and Holocene near-shore and beach deposits (Santanach et al., 2011).

**Monday, March 31**

Parc Güell is located in the hills north of downtown Barcelona. The park is currently a garden with architectural features designed by Antoni Gaudí and constructed between 1900 and 1914. Originally, Count Eusebi Güell intended it to be a housing project inspired by the English garden city movement. Unfortunately, there was little interest in this upper-class housing development, only two houses were built, and it eventually became a municipal park. The site is located on outcrops of Silurian to Devonian limestones and shales and Silurian dark shales and chert. The dark shales and phyllite can be seen throughout the park, forming a crumbly, but relatively resistant hillside.

**Tuesday, April 1st (April Fool’s Day)**

Montserrat

Montserrat Mountain (“serrated mountain”) is approximately 50 kms. north of Barcelona. It is a ridge of Paleocene conglomerate that rises in the Pre-Coastal mountain range adjacent to the Llobregat River. The mountain is approximately 8 kms long and 5 kms wide, with the highest peak, St. Jeroni, elevating to 1,236 meters. The greatest relief is along the northern margin, near the towns of Marganeill and Monistrol. The rounded peaks of conglomerate create a very unique character. We took a train to Montserrat, a cable car (Aeri),
and a funicular up toward Sant Joan. We then hiked over to St. Jeroni and back down to the monastery.

Paleogene transpression (50 Ma) uplifted the Catalan-Balearic massif, and created the Ebro foreland basin and an inland sea. Fluvial drainage systems emanating from the Catalan Coastal Range, to the south, deposited sediment in two fan-deltas systems along the margin of the sea (Montserrat and Sant Llorenc del Munt system). The fan-deltas prograded into the Ebro foreland where rivers from the uplands drained through the Catalan Coastal Range front, possibly through tear faults. These systems deposited over 1,000 meters of coarse-grained fan deposits. Starting in the Middle Eocene and continuing for 4.4 my, proximal debris-flow, sheetflood and distal fluvial deposits created and maintained a fan surface at or above sea level despite rapid subsidence of the foreland. Palynology suggests a warm, humid climate existed during this interval (López-Blanco et al., 2000).

The Ebro foreland basin was wedged between the Pyrenees, Iberian Cordillera, and the Catalan Coastal Range.

The texture of the conglomerates exposed at Montserrat clearly indicate a very high-energy depositional setting. The conglomerate that we observed on our walk consisted of round to well-round clasts up to 50 cm in diameter in a sandy matrix but clasts of up to 1 meter have been reported. The majority of the clasts appear to be carbonate (Mesozoic carbonates eroded from the uplands to the south?). We saw some occasional sand beds but they were limited in extent and very coarse. It is a very beautiful conglomerate. The monastery utilized the conglomerate in its construction; several of the columns are made of polished conglomerate.

Tectonic uplift of the Pre-Coastal mountains during the Alpine orogeny (Oligocene) and erosion of the less resistant strata produced the Montserrat range. The compression (and subsequent extension) generated a vertical joint pattern that enhanced erosion and produced to the
column-shaped pinnacles. Dissolution of the carbonate cement contributed to the weathering of the pinnacles.

Monday, April 7th
Marseille Calanques

The marine transgression associated with the end of the last glaciation inundated karst topography that had developed in the Urgonian limestone between Marseille and Cassis, France. These carbonates are part of the Massif des Calanques which is characterized by deep, narrow coastal drainages referred to as calanques. In some cases the calanques formed as a result of roof-collapse that produced sinkholes. Although calanques are found elsewhere in the Mediterranean, the calanques of the Marseille region are the most spectacular and extensive. Some of the calanques may have initially developed during the Messinian salinity crisis (5.96 to 5.32 Ma) when the Mediterranean Sea was isolated from the Atlantic and evaporation dropped sea level 500 meter below Atlantic sea level. This drop in sea level resulted in the accumulation of evaporates on the Mediterranean abyssal plains and promoted deep erosion by rivers that drained into the basin. The geomorphology of the continental shelf indicates paleo-shoreline at depths of 36, 50, 90, and 100 meters below present sea level. Radiocarbon dating of the 100 meter shoreline has given an age of 13,250 ybp (Collina-Girard, 1996). The Cosquer Cave is found at the 37 meter level below Cap Morgiou. This cave contains
Paleolithic paintings and engravings that have been dated and indicate two periods of occupation, 27,000 and 19,500 ybp (Collina-Girard, 1996).

As seen in the adjacent map, the fracture pattern has a distinct NW-SE and SW-NE fracture orientation. A national park (Parc Nationale des Calanques) was established in 2012 to protect the limestone calanques of the Massif des Calanques.

We took a ferry from Marceille, past Château d’If to Frioul Harbour on Ratonneau Island and walked across the causeway to Poméguès Island to see the karst features. It is about a 4 km hike down and back the fortified island that is now part of the national park. The white carbonate is gently folded and contains occasional fractures that contain calcite crystals and heavy brown iron oxide deposits. Fluids moving through the rock remobilized the carbonate and calcite crystals grew in the open voids created along some of the fracture zones. Similarly, iron was remobilized and precipitated in these fractures as well.

**Tuesday, April 8th**

Monaco is located in the southern extension of the Alps, uplift and deformation during the Alpine orogeny resulted in the steep topography of the region. Carbonate rocks can be seen in the hills above Monaco.
Corsica is made up of rocks of European affinity to the west (Hercynian Corsica) and “Internal Units” characterized by oceanic substrate and Jurassic ophiolites (Alpine Corsica) to the east. Hercynian Corsica consists of Upper Carboniferous to Permian (late Variscan orogeny) granitoids intruded into Precambrian and Paleozoic country rock. In the west some uneroded Mesozoic to Upper Eocene sediments remain. The ophiolite sequences are interpreted as part of the western Tethys ocean basin that had developed the Europe/Corsica and the Adria continental margin. Late Cretaceous convergence in the Tethys basin created intraoceainic subduction and, eventually, continental collision between Europe/Corsica and Adria continental plates (Bortolotti et al., 2004). During compression, the ocean crust was deformed and HP-LT (11 kbar, 400 °C) metamorphosed at depth in the subduction zone (more weakly metamorphism is found in the Apennines). The ophiolites are found in the “Schistes lustrés” and are associated with units derived from the continental margin (Bortolotti et al., 2004; Lahondere and Guerrot, 1997).

The granite that makes up the mountains of western, Hercynian Corsica predate the Alpine orogeny, they were the product of a much earlier continental collision and have ages that range from 315 to 280 Ma. Paleozoic (Precambrian?) rocks are strongly deformed and metamorphosed by the Hercynian orogeny. The metamorphic gradient decreases to the northeast in Corsica and the structural trends are northwest-southeast. In southern Corsica there exists
amphibolite facies which are sometimes strongly migmatised (Sartori and others).

Thrust nappes developed in the Alpine Corsica. These unmetamorphosed allochthonous units overlay the highly strained metamorphic complex. Since 33 Ma, the region has experienced extension, the Western Mediterranean sea began to open and, as the Sardinia-Corsica block was displaced and rotated, the nappes slid down a major fault zone that runs across Corisica from the north to the southeast. Currently, the nappes can be found east of the fault zone and the underlying granite moved up on the west side (Fournier et al., 1991).

Post-Alpine deposition occurred in several basins in Corsica: the Bonifacio, Corte, and Nebbio basins as well as accumulations in the eastern coastal plain. Late Oligocene to Miocene north-trending rifting created extensional basins. Pre-rift deposition (late Eocene – early Oligocene) are continental; syn-rift sequences grade from continental to marine facies and include calc-alkaline volcansics due to the westward-dipping subduction zone that developed as the Provencal basin opened. Post-rift deposits include dirty limestones, lagoonal and lacustrine facies.

Spheroidal and Tafoni can be found on the granites in the hills south of Calvi. Spheroidal weathering is… Tafoni weathering is found throughout the world in all types of climates but the process is not well understood. Wind erosion, microclimate temperature variations and salt crystallization have been suggested to cause these large-scale cavernous features (Mustoe, 1982; Hume, 1925).
Monday, April 14th
Mount Vesuvius

After a tram, train and bus ride, we climbed to the Mt. Vesuvius (Monte Vesuvio) crater. Vesuvius is a stratovolcano on the west margin of the Gulf of Naples and part of the Campanian volcanic arc. The Adriatic Sea is bound on both sides by subduction zones, on the east, the African plate is being subducted beneath the Eurasian plate and to the west it is being subducted beneath the Italian Peninsula (Eurasian plate), creating the Campanian volcanic arc. This volcanic arc includes all the active volcanoes of Italy: Campi Flegrei (Phlegraean Fields), Vesuvius, Stromboli, Panarea, Vulcano, Etna and Campi Flegrei. It appears that beneath Vesuvius the descending slab is detached from the upper slab, creating a “slab window”. This may explain the unique composition of the lavas of Vesuvius. Convergence on both sides of the Adriatic Sea is closing the basin.

Somma-Vesuvius consists of two cones: a younger large cone (Gran Cono) that is 1,281 meters high, partially enclosed by the flanks of an earlier collapsed cone, Monte Somma, 1,149 meters in elevation. Vesuvius and Somma are separated by a Valle del Gigante (Atrio del Cavallo on the west and Valle dell’Inferno on the east). The Vesuvius crater is 650 meters in diameter (this makes it a crater, not a caldera) and 230 meters deep. The volcano is generally referred to as Somma-Vesuvius. Somma-Vesuvius is built on a 34,000 year old ignimbrite generated by the Campi Flegrei complex (northwest of Naples). There is fumarole and degassing activity within the crater, it is noticeable on the west side as you walk the crater rim.

The composition of the lava is variable, but in general it is silica-undersaturated and enriched in potassium, ranging from slightly undersaturated basalt and trachyte to highly undersaturated tephrite to phonolite; phonolite was produced in some of the larger
eruptions (Di Renzo et al., 2007). The interplinian magma are less evolved than the plinian lavas. It has been suggested that periodic influx of basic magma from depth (12 kms) would mix with the differentiated shallow magma (5 kms) causing a plinian eruption (Hermes and Cornell, 1981; Cortini et al., 1985; Barberi and Leoni, 1980). Continued influx of basaltic magma may prevent differentiation and interplinian eruptions occur. As higher-viscosity phonolitic magmas develop, interplinian eruptions cease until the cycle starts again. Guidoboni and Boschi (2006) identified a eruptive trend: a plinian eruption followed by interplinian activity and a period of quiescence.

Somma-Vesuvius started forming approximately 25,000 ybp with a series of lava flows and minor pyroclastic eruptions. The eruptive style changed around 19,000 ybp, producing explosive plinian eruptions. These include the Basal Pumice (Pomici di Base) eruption 18,300 ybp that produced the original Somma crater, the Green Pumice (Pomici Verdoline) eruption 16,000 ybp, the Mercato eruption (Pomici de Mercato) eruption 8,000 ybp, the Avellio eruption (Pomici de Avellino) 3,550 ybp, as well as eruptions in 79 AD, 472 AD and 1631 eruption. The Avellino eruption vent was approximately 2 kms west of the present crater which destroyed several Bronze-age towns on the western flanks of Somma-Vesuvius.

Between plinian eruptions there is a history of minor eruptive events ranging from subplinian to more or less violent strombolian emissions that often involve lava flows (Guidoboni and Boschi, 2006). Some of the eruptions where phreatomagmatic. Numerous strombolian-phreatomagmatic events and extensive lava emissins formed the summit cone of Vesuvius between 472 and 1631. The collapse of the Somma caldera is subject to debate. Some believe the collapse was related to the Avellino plinian eruption (3,550 ybp) others argue the cone was destroyed earlier (Rolandi, 1998). The Vesuvius cone appears to have developed through a series of non-plinian events since the 1631 eruption (Rolandi, 1998). An eruption in 79 AD made Vesuvius notable because of the destruction of the Roman cities of Herculaneum and Pompeii. This eruption lasted two days; pyroclastics were ejected at an estimated rate of up to 1.5 million tons/second. Pyroclastics (ash) blanketed the area during the first day and most people evacuated the towns nearby. When the high-altitude eruption columns (15 – 30 kms) collapsed during the second day, pyroclastic flows (pyroclastic density currents) traveled through Herculaneum and Pompeii and killing an estimated 16,000 people.
The eruption alternated between plinian and pelean six times and the 3rd and 4th pyroclastic surges are believed to have destroyed Pompeii. These pyroclastic flows altered the landscape, including the coastline.

Vesuvius has erupted several dozen times since then and is the only volcano on mainland Italy that has erupted in the past 100 years. An eruption in April, 1906 killed over 100 people and produced voluminous lava flows. The most recent eruption, in March, 1944, destroyed several villages.

It is a Decade Volcano because over 3 million people live near the mountain and it has a history of plinian eruptions. The emergency plan for Somma-Vesuvius is based on an eruption the size of the 1631 event. A 7 km. radius around the present crater has the potential for pyroclastic flows and tephra fallout would cover a much greater area down-wind (south and east). A tephra accumulation of greater than 100 kg/m² may result in roof collapse. The emergency evacuation plan assumes a 14-20 day forecast to move up to 600,000 people from harm’s way in 7 days. They are actively trying to reduce the number of people that live in the red zone and have established a national park that encompasses the upper flanks of Somma-Vesuvius (8,482 hectares). The volcano is currently monitored by the Osservatorio Vesuvio in Naples using a network of seismic, gravimetric, GPS-based geodetic, local surveys and fumarole chemical analysis.
Thursday, April 10th
Messina Earthquake

A magnitude 7.1 earthquake, that lasted approximately 40 seconds, leveled the cities of Messina and Reggio Calabria on December 28, 1908. It was the most deadly earthquake in the history of Europe; the earthquake and subsequent tsunami took an estimated 100,000 to 123,000 lives. The epicenter has been modeled between Sicily and Calabria, on the Italian mainland based on the orientation of toppled monuments (Omori, 1909). Motion along the fault appears to have a normal, generated by vertical stress induced by the subducting African plate. Subsidence occurred on both sides of the Straits of Messina. The tsunami was either caused by motion along the fault or a large submarine landslide triggered by the earthquake.

Buildings were not constructed to withstand earthquakes and their heavy roofs and weak foundations resulted in the destruction of 91% of the structures in Messina. The damage was exacerbated by a series of significant earthquakes between 1894 and 1908 with magnitudes greater than 6; these events weakened the poorly constructed buildings. Fire contributed to the city’s destruction (Pino et al., 2009). The tsunami had a mean run-up 5 meters with a maximum of 12 meters south of Messina and Reggio Calabria (Pino et al., 2009).

Messina and Reggio Calabria were, once again, heavily damaged in July, 1943 as the Allies, who landed in southern Sicily, drove the Germans and Italians from Messina to Reggio Calabria. In September the Allies invaded the Italian mainland at Salerno.

Mount Etna
Mount Etna is a large stratovolcano with a summit at 3,329 meters (10,922 ft) that is unusually active. Due to this activity and large population center in Catania (315,000 residents), Etna has been designated a Decade Volcano (as is Vesuvius). Etna began erupting tholeiitic basalts approximately 500,000 years ago off the coast of ancient Sicily. Starting 170 ka, mafic alkaline magma (pigeonitic tholeiites, alkali basalts and trachybasalts) was emitted from several vents to the southwest side of Etna referred to as the “ancient alkali centers” (Romano, 1982; Gillot et al., 1994). This early phase is referred to as “pre-Etnean”. The volcanic vents moved to various positions on Etna and over the course of time eruptions of pyroclastics and lavas built the stratovolcano.
The plate tectonic setting of Etna is interesting. The volcano is part of the Calabrian volcanic arc, the product of subduction of the Adriatic (Ionian) Sea beneath the Eurasian content. Some consider the Adriatic part of the African plate while others interpret a transform margin separating the African plate with the Adriatic micro-plate. Others have suggested that Etna is a hot spot in order to explain the high lava production and mafic magmas (Tanguy et al., 1997; Schiano et al., 2001). A final theory is that Etna is part of a complex rifting complex with the volcano located at the intersection of several major regional fault systems (Behncke, 2001). This recent hypothesis would make Etna genetically distinct from the volcanos of the Aeolian Islands. Factors that support an extensional regime include: (1) movement between the Malta-Sicilian block and the Ionian basin (Geronimo et al., 1978; Gillot et al., 1994), (2) formation of a graben in the Catania plain, (3) intersection of Malta Escarpment and Messina-Giardini fault zone (McGuire et al., 1997), extension along the east-facing normal fault in the Siculo-Calabrian rift, and (7) rollback of the subducted oceanic lithosphere below the Tyrrhenian Sea (Doglioni et al., 2001).

Between 35,000 and 15,000 ybp Etna entered an explosive phase that generated large pyroclastic flows and left extensive ignimbrite deposits.

Approximately 8,000 years ago the eastern flank of Etna experienced a catastrophic collapse that produced a scarp known as Valle del Bove (Valley of the Ox) and may have generated a tsunami that inundated parts of the eastern Mediterranean. The collapse created a depression approximately 5 kms by 7 kms. The sector collapse generated a debris avalanche (Calvari et al., 1998). A second collapse, around 3500 ybp, once again changed the shape of Valle del Bove. Another collapse occurred around 2,000 years ago and formed Piano Caldera; this collapse has been associated with a plinian eruption in 122 BC (Coltelli et al., 1998). This caldera has been subsequently filled with lava but can still be seen as a distinct break in the slope of the mountain near the base of the present summit cone (2900 m).

Although the volcanic activity of Etna has been nearly continuous, the eruptions exhibit variations in location and style. There are several historic craters near the summit: Northeast Crater, Voragine, Bocca Nuova, and Southeast Crater Complex, and over 300 vents on the flanks that range from small holes to craters 100’s of meters in diameter. Because of the size of Etna, the summit eruptions are generally not hazardous but the flank eruptions can be very close to populated areas.

Although most of Etna’s eruptions have been non-violent, there have been some plinian events; in 122 BC a violent plinian eruption resulted in voluminous tephra that cause roofs to collapse in
Catania (the Roman government waived taxes for the residents of Catania for 10 years to allow them to rebuild).

The list of documented eruptions of Etna is greater than any other volcano in the world, going back to 1600 BC. An eruption in 1669 generated a lava flow that destroyed at least 10 villages on the southern flank of Etna. This lava flow extended for 17.3 kms, fed by a lava tube, and reached Catania but was diverted by the city walls into the sea and filling the harbor south of the city. The one place the lava flow breached the city wall, it stopped short of the Benedictine monastery but didn’t reach the center of town. This eruption produced approximately 1 km3 of lava and lasted about four months (Behncke et al., 2005).

Since 1600 there have been at least 60 flank eruptions and countless summit events. Since 2000 there have been four flank eruptions (2001, 2002-3, 2004-5, 2008-9) and two summit eruptions (2006, 2007-8, 2012-13). When we drove up toward Etna, fresh coarse black ash could be seen on roofs and on the shoulders of the road north of Catania.

An eruption in 1928 on Etna’s northeastern flank generated a lava flow that inundated the village of Mascali. The eruption started high on the flank and new eruptive fissures opened at lower elevations with the third, and most destructive, opening just 1,200 meters above sea level (Ripe della Naca). Lava from an eruption in 1971 destroyed the Etna Observatory and the first cable-car system; it also threatened several small villages on Etna’s east flank. An eruption in March 1981 the town of Randazzo on the volcano’s northwestern flank was threatened by a lava flow. Eruptions in 1991-93 threatened the town of Zafferana but diversion efforts diverted disaster. The diversion efforts included construction of a earth barrier to slow the flow and when that wasn’t successful, explosives were used to close the 7 km. feeder lava tube and divert the flow into an artificial diversion channel. A lava flow in July, 2011 was also diverted successfully when it threatened the Sapienza Refuge.

Hazards associated with Etna include pyroclastics (tephra), primarily from summit eruptions, and lava flows, primarily from flank events. Scollo and others (2013) assessed the impact of tephra fallout from short and extended eruptions. They found that, due to prevailing wind directions, the eastern flanks of Etna are most affected by tephra fallout. Infrastructure and agriculture are impacted by this pyroclastic deposition. Ash can also produce chronic health effects such as silicosis and chronic pulmonary diseases (Horwell and Baxter, 2006).

Lava flow probability maps for summit eruptions and flank eruptions based on models by Negro and others (2013).
Monday, April 25th
Venice, Italy

Venice is sinking; a combination of subsidence and sea level rise is causing the city to flood. This is a similar situation that New Orleans faces, except the intricate relationship between the structures and the lagoon means is no way of building a system of dikes around the city of Venice. In both Venice and New Orleans, subsidence and sea level rise will not be reversed leaving a continuous battle that will not be won.

Unlike New Orleans, Venice rarely experiences strong storm surges but on November 3rd, 1966 a violent storm pushed a surge into the city and increased the water level to up to nearly two meters above mean sea level. This event was a wake up call for the city and those throughout the world that sought to preserve the unique architecture of Venice. UNESCO and the Italian government identified 100 structures that needed stabilization and remediation. One of the most detrimental aspects of structural subsidence is that seawater now rises above the protective Istrian stone (impermeable marble) foundations of the buildings and saturates the brick. The salt is absorbed by the brick, wicks upward, and degrades it.

Venice is constructed on a group of over 100 small islands, both natural and artificial within the Venice lagoon. The lagoon is approximately 550 km² in area and averages one meter in depth. Structures are built on millions of piles that have been driven into the fine, compressed sediments of the underlying lagoon (10,000 piling support Rialto Bridge alone). The wood piles are between 2 and 3 meters long and 25 to 30 cms. in diameter (). Since they never come into contact with air (oxygen), they are generally unsceptible to rot.

The region has been experiencing subsidence since Venice was first founded in the 9th century A.D. at a rate of approximately 10 centimeters per 100 years. This “natural” tectonic subsidence is because the region is within the foreland between the Alps and Apennine mountains (Scearc, 2007; Carminati et al., 2005). During the past century that rate has increased significantly to 25 cms. (Spencer et al., 2005; Frassetto, 2005). Ground water mining by communities on the mainland and extraction of gas has been blamed for the increased subsidence rate. The ground water and gas extraction was halted in the 1970’s
but subsidence continues. Seasonal floods, the acque alte, are caused by high tides, low-pressure systems (most frequent in the late fall), and winds that surge water across the lagoon. These events flood St. Mark’s Square and other low-lying areas of Venice. In 1900, the frequency of these floods were approximately 6 per year, now they occur 60 times a year. When we walked through St. Mark’s Square in the evening, duckboard walkways were installed and water was surging out of the drains, filling the square.

In addition to the city subsiding, the lagoon is being flooded as well. The ecosystem that supported sea grass that stabilized the sediment is becoming deeper and saltier, behaving more and more like open water. Originally, there was an substantial influx of freshwater into the Venice lagoon but, over time, the salinity has increased with increased exchange with the Adriatic; today the lagoon salinity level is near that of the sea (Scearce, 2007).

Channels that have been dredged for shipping are also accumulating sediment from the lagoon. Water quality, particularly near the city, is in poor shape. Eutrophication, chemical toxins, heavy metals and endocrine disrupters have affected water quality within the lagoon (Fletcher et al., 2005; Zirino et al., 2005).

The Centro Previsioni e Segnalazioni Maree (CPSM) provides sea level forecast and warning system for Venice (Scearce, 2007). They utilize observations of sea level and meteorology, short-term sea level forecasts, tidal data and issues high water level event alerts.

Three inlets exist between the Adriatic Sea and Venice lagoon, Lido Inlet, Malamocco Inlet, and Chioggia Inlet (from north to south). One proposed solution was to construct a mobile barrier between the lagoon and the Adriatic, the Modulo Sperimentale Elettromeccanico (MOSE) Gates. The estimated cost of this project is in the 3.5 to 4.3 billion euro range. The gates are intended to remain open the majority of the time but during high water events the gates would be raised, damming the inlets and keeping flood waters...
from entering the lagoon. This engineered solution is expected to buy Venice 100-200 years of protection (Scearce, 2007).

Another proposal was to convert the lagoon into a freshwater lake and pump the water out to the sea. The Dutch have done this with part of the Eastern Scheldt estuary. Another proposal is to pump water into the subsurface in order to reverse subsidence but this has been found to be technically unfeasible (Cocks, 2005).
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Spencer et al., 2005; Frassetto, 2005).


Hermes and Cornell, 1981; Cortini et al., 1985; Barberi and Leoni, 1980).