Summary

A field excursion involving 8 students and 1 instructor was conducted in the northern part of Tanzania, along the East Africa Rift System between 03rd and 06 March, 2017. The aim of the excursion was to study volcanic processes, features, hazards and monitoring techniques. The learning modality was in form of lectures and field trips. The important volcanic aspects covered were the overview on the East Africa Rift System, Overview on volcanic types, style and associated geomorphological features, timing of the major tectonic movements and volcanism in relation to the East Africa Rift Valley, Lake Natron fault boundary system and deposits. Others were the Engaresero hominid foot prints showing the oldest, tuff ring craters, sink pits and cones/hills along the East Africa Rift system. Field observation has indicated that the rift system is stretching rapidly in geological sense as evidenced by very recent sink pits (2008-present) and various cones which are cut by the rift system. This is also supported by high frequencies of volcanic eruptions in the recent years (eg. Mt. Oldoinyo Lengai). However, such frequencies do not match with the government effort in ether awareness or protection of people and their properties from volcanic hazards. More efforts, particularly governmental efforts are still needed in creation of awareness to local people living in volcanic risk zones. Alternation of sediments and volcanic materials in various parts of the rifts suggest many episodes of volcanism and tectonics associated with changes in climatic conditions. Most of the available data records recent events. Therefore, proper dating of materials along the rift is recommended for magma evolution and proper reconstruction of volcanic activities along the rift. The current climatic condition is highly favoring formation of trona/magadi in shallow lakes along the system (e.g. Natron). Industrialization on magadi focusing on proper utilization of such materials as a source of income to local people and the country at large is recommended. As the rift system is characterized by different geological formations, more research on materials sciences, energy and minerals are recommended.

Introduction

East Africa Rift System Overview

A series of rift valleys is the typical geomorphological character of the Eastern Africa. These valleys are generalized as the East Africa Rift Valleys (EARV). The EARV extends to about 4000 km, from north (Ethiopia) to south (Mozambique). It consists of adjacent elongated depressions or tectonic basins, generally bordered by uplifted shoulders. The whole system is fault bounded, the lower, subsided part is called “graben”, the high bordering relief is called “horst”. The graben is well developed in Kenya and in Tanzania, a half-graben is well pronounced particularly close to the border with Kenya. The higher borders of the rifts are usually parallel mountains, plateaus, sometimes volcanic massifs. The rift valley has an average width of 50 km and an average bounding scarp height of 600 m, important differences exist depending on the section. For instance, in Kenya, the Gregory Rift is about 100 km wide and part of its eastern shoulder, the Aberdare Mountains, has an elevation of about 2000 m above the rift floor. The EARV can be subdivided into several sections:

- The Ethiopian Rift extends from the Afar Triangle to Lake Turkana in Kenya
- The Eastern Rift, or Gregory Rift (named after the American geologist J.W.Gregory) in Kenya and Tanzania. In Tanzania it is called the Northern Tanzanian Divergence and includes the branch in which Lake Eyasi lies
- The Western rift includes the segment from Lake Albert up to Lake Malawi (or Nyassa), including the Rukwa section
- The Malawi Rift includes Lake Malawi and the Shire Valley, also the Urema basin(Mozambique) and Luangwa Valley (Zambia)

Rifts are not randomly located, they are concentrated on specific weak zones. The East African rift valley resulted from the east-west spreading of two micro plates, the Somalian plate moving away from the Nubian plate. Analog models suggest that pre-existing weaknesses within the continental lithosphere may localize deformations. In both branches of the rift valley there is a relation between the basement tectonics and the rifting, as normal faults are localized in mobile belts (Kibaran, Ubendian, Ruwenzori, Mozambique belts) along the edges of the Tanzanian craton. The surrounding mobile belts are to some extent weaker than the craton (Petit and Ebinger, 2000). The location depends on the activation of pre-existing weaknesses in the lithosphere. The rift faults follow these weaknesses with respect to the direction of the extension. Orthogonal rifting, a system where the rift-border faults and intra-rift faults develop perpendicular to the extension vector, gives rise to long extension-orthogonal boundary faults with pronounced subsidence. Oblique rifting, a system where the orientation of the border faults is controlled by the geometry of the underlying basement stretching zone, and intra-rift faults form orthogonal to the extension vector, and results in a general en-echelon arrangement of faults and basins

with less subsidence. Magmatism and volcanic activity may occur along the boundary faults, transfer zones and parts of the rift shoulders.

Referring to rifting, Chorowitz and others (1987) describe several stages of rift formation:

- Pre-rift stage: dense fracturing, with strike-slip faults while a shallow, wide depression forms.
- Initial rifting stage: Occurrence of tholeiitic volcanism with oblique-slip faults, bounding tilted blocks, but uplift of rift shoulders is not very pronounced.
- Typical rift formation stage: normal faults bordering the main tilted blocks with subsidence of the rift floor and important uplift of shoulders.
- Advanced rifting stage: important magmatic intrusions along the rift axis. This stage corresponds with the initial formation of oceans. These stages correspond today with several sections of Eastern Africa.

The developing extension of the rift system leads to a change in deformation style, from long boundary faults at rift-margins to dense fault swarms affecting rift floor where magmatic activity is concentrated. The lithospheric crust is thinned and weakened by magma intrusion which will lead to the rupture of the continental crust as the final result. In these areas, magmatic segments act like slow-spreading mid-ocean ridges, developing within a lithosphere which is transitional between continental and oceanic (This is happening in Ethiopia).

The EARV is characterized by the different stages of rifting. The African rift system developed southwards from the Afar area in Ethiopia, and developed around the Tanzania craton in two branches. It is a system of fault-bounded elongated basins, with or without visible magmatism, which run N-S from the Afar depression to Mozambique. The eastern branch (the Ethiopian and Kenyan rifts and the Northern Tanzanian Divergence) displays important volumes of magmatism but a very reduced seismic activity, with focal depths of maximum 15 km. The western branch has only little volcanic activity but frequent and large earthquakes, with deep events of more than 30 km (Yang and Chen, 2010). The beginning of volcanism and probably the extension activity was considered older in the eastern branch than in the western branch, but a recent study by Roberts and others (2012) puts the rift timing for both branches in the same time frame. Their studies are based on detrital zircon geochronology, tephro- and magnetostratigraphy and analysis of past river flows in sedimentary rocks in the Rukwa Basin.

The eastern branch of the Rift Valley mainly follows the north-south trend of the Mozambique fold belt. The eastern rift system is easier to follow in Kenya, with better developed sub-parallel structures, whereas in Tanzania the rifting only enters little and is blocked by the Tanzanian craton and splayed out eastwards to form the Northern Tanzanian Divergence (Dawson, 2008). Unlike the western rift, the Gregory rift is characterized by more volcanic activities. Magmatism started in Ethiopia about 30-40 Ma, continuing southward into Kenya around 12 Ma, and arriving in Northern Tanzania around 8 Ma (Dawson

1992, 2008). In Ethiopia and Kenya, the larger volcanic areas, the early volcanic activity and its eruption materials are largely covered by huge amounts of volcanic rocks. In Tanzania, a small volcanic province, more recent volcanic rocks with mantle xenoliths have been studied. These provided evidences for a complex mantle history and distinguished various volcanic features. During the rifting processes, sedimentary basins which are important for dating and understanding the rift valley system were formed. The end of Miocene tectonic events is one among the tectonic processes associated with major deposition of sediments

2.0 Volcanic Types and Styles

The common known volcanic types are shield volcano, stratovolcano and cinder cones. All of these are commonly known along the East Africa Rift system.

Shield volcano: These are volcanoes characterized by low viscosity and low silica materials, particularly of basaltic formation. Due to low viscosity, such volcanoes cover a wide area at low angle (gentle slope generally less than 10 degrees for the created volcanic cones). Shield volcanoes are polygenetic central volcanoes, where the activity is always centered on the same vent. Ngorongoro, Olmoti, Kitumbeine, Gelai and Shira volcanoes are the best examples of shield volcanoes in Northern Tanzania. They were formed during and after a first series of major tectonic movements, when huge amounts of basaltic lava filled, and overspilled the late Tertiary depression. Out of the EAR system, the best examples of shield volcanoes are the islands of Hawaii, which form a linear chain of volcanoes formed by a hot spot, and moving away from it by the NW directed movement of the Pacific Plate. Mauna Loa is the biggest volcano in the world (slightly higher than Mt. Everest if we consider the height from its base under the ocean). It rises with a 10% medium slope, creating a massive volcano.

Stratovolcano: These are sometimes known as composite volcanoes. They are polygenetic central volcanoes with a very pronounced conical structure, with lower flanks slightly inclined but with upper slopes of angles of 30-35%. They are formed by alternation of lava and pyroclastic materials sometimes with materials eroded from the upper slopes. Their steepness is sometimes attributed to the alternation of resistant (lava) and weak (pyroclastic) materials where the lava protects pyroclastic materials from erosion. Intermediate composition (andesitic) is common. Well known examples of such volcanoes in northern Tanzania are Essimingor, Kerimasi and Mt. Meru volcanoes. Globally, huge mountains are classified in this volcanic type. Such mountains are mostly distributed along two major belts, at plate margins: the circum-Pacific Belt,or “Ring of Fire” (ex. Mt. Fuji) and the Mediterranean Belt (ex. Mt .Vesuvius). Dormant periods of these volcanoes can be tens, hundreds or even thousands of years in which pressure builds up under the volcanoes, and usually result in very explosive and disastrous eruptions.

Cinder cone: These are monogenetic cones produced in a single eruption phase. These are of different chemical composition and therefore can be classified according to the texture of the produced

materials. For instance, scoria cones, ash cones, pyroclastic cones are common terms. Most of these materials land close to a central vent and build up a cone with typical slopes of up to 30% (33% is the gravitational maximum for this loose material). Cinder cones are small and seldom exceed 500 m of height. Geologically, their life span is short. They are susceptible to erosion due to the steepness of the cones and the nature of materials .

Spatter cone: A spatter cone is a small, steep-sided, monogenetic volcanic cone built up by tephra blown out as strips of fluid basaltic lavas, and melt together after deposition. They occur on the slopes of shield volcanoes and their heights are usually between 5 and 20 m.

Volcanic dome: These are associated with viscous magmatic materials, rich in silica. They are steep-sided dome- or spine-shaped mound of viscous lava, creating a growing mass of volcanic rock above a volcanic vent. The domes can be chemically characterized by rhyolite to andesitic composition if there is crystallization of minerals. It can be obsidian (volcanic glass) if minerals do not crystallize. The domes grow under more injection of these lavas from the deeper parts. As this structure acts like a plug for gases, which cannot escape, the pressure inside will grow till a major explosion takes place. These domes have been associated with some of the most destructive volcanic activity worldwide.

Calderas: These are large depressions resulting primarily from a partial collapse of the volcanic edifice. Like pit craters they are the result of this collapsing, but they are larger, minimum 1 km in diameter. They form when a volcano's magma chamber is emptied and through the weight of the overlying rock, the volcano's top collapses, forming the caldera, which usually looks like a very “large crater”, but for the very large calderas, their shapes and alignment can reflect the structural features of the underlying crust. Calderas have been classified in various types, all resulting from collapses, but the biggest distinction is their composition -mafic or felsic- which results into different form and eruption style. The largest known caldera is the Toba caldera in Sumatra (100 km long, 30 km wide), an elliptic shaped, felsic caldera, related to a tectonic structure, called the Semangka rift zone. In Northern Tanzania, in the crater Highlands, several mafic calderas were formed, like those of Ngorongoro or Olmoti volcanoes, or further east the one of Mt. Meru, contrasting as one of the more felsic type. A caldera is not an end stage of volcanic activity. Most calderas had many eruptions over tens or hundreds of thousands of years. During the dormant periods, the magma chamber replenishes, resulting in an uplift of the caldera floor and small cones or domes might erupt, fed by the new magma input. As an example, we note Mt. Meru's caldera showing a post-caldera forming cinder cone in its center.

Tuff rings: These are formed when the magma interacts with water under high-energy eruptions by which debris is accumulated around a vent (with a water-to-magma rate of 0.1 to 1). This explosivity results in a spreading of volcanic material and creates a crater which opens, or partially opens, in the rock. There is a tenet by which CO_2, instead of water, could be the erupting agent for the explosive mechanism (Sherrod and others, 2013) Maars are similar craters, more or less circular, usually filled with

water, with its base under the original surface of the land. Tuff rings and maars are common along the EAR system (eg. Lake Duluti is a typical maar).

Hornitos: Hornitos are small cones or chimneys of several meters high, consisting of lava spatter, and found on the surface of pahoehoe flows. The building up is done by lava spatter thrown out of the underlying flow by the explosive release of magma gases or by explosivity created by steam from groundwater beneath the flow. Hornito building is typical for a volcano like Ol'Doynio Lengai.

Flood basalt: Flood basalts are not volcanoes .Originating from an upper mantle melt source, it seems to have no shallow storage. The basalts have a very low silica content (very low viscosity) , meaning they are very fluid . They spread out in large sheets which can cover hundreds of square kilometers. For the case of rift systems, e.g. EAR, when a mantle plume underlies a rift, the divergent movement of the overlying rift can form the head of the plume fissure, resulting in the release of huge masses of partially molten rock which the plume contains. The plume head shows a higher temperature than the normal asthenosphere, indicating that more partial melting can occur in it and very large quantities of magma can be released. The bigger flood basalt areas are called LIPs (large igneous provinces). Examples are the Columbia River Plateau in Washington State, USA, the Deccan Traps in India, or the Ethiopian Plateau, part of the East African Rift.

In human history, no LIP has been witnessed. Basalts can also take a columnar shape and a known good example is the columns of Giant's Causeway, Ireland.

Volcanic eruption types

Basically, there are two types of volcanic eruptions being controlled by gas content, chemical composition/viscosity and the quantity of volcanic materials. These are ***effusive and explosive eruptions .***

An effusive eruption mostly happens when the degassed magma overflows from a crater or a vent and form a lava flow. Lava flows can flow for hundreds of kilometers if they are composed of low viscosity basalt. Sometimes, a lava lake can form around a vent in an effusive eruption. Sometimes a cinder cone can be formed around a vent when lava sprays up in fountains. Underwater eruptions usually produce effusive eruptions as the pressure of the water prevents or limits the liberation and expansion of gases necessary for the explosive activity. The main type of effusive eruption is the Hawaiian one.

An explosive eruption (or pyroclastic eruption) is the eruption associated with very high pressure where some of the magmatic materials are thrown into the atmosphere and fall on the ground under the influence of gravity. Pyroclastic materials and gases are common in such types of eruptions. It can lead to the complete destruction of a magma chamber and expel in the atmosphere fragments of this magma at high temperature. This explosive activity results from the violent expansion of the gases in the liquid

magma. With the rising of the magma the pressure descends and gas is released. Explosive eruptions can generate clouds ranging from a couple of feet rising in the air to 50 km high.

Examples of explosive eruptions are the Strombolian, Vulcanian, Plinian and subplinian ones.

3.0 Natron basin fault system

One of the observed features during the excursion is the Natron basin fault system. The trend of this system is mostly N-S, and from the Kenyan border to the south end of Lake Natron several segments can be distinguished. The Sanjan fault is the most westerly of these faults, and between this section and the others lies the Moinik Valley, with extensive sedimentary deposits of the proto-Lake Natron. The height of the faults go down from c.1200 m at Ol'Doynio Sambu to nearly zero meters north of Engare Sero, and up again to 600 m south of Lake Natron. South of Ol'Doynio Lengai, it turns SE for a while until Kerimasi volcano, where it disappears under the volcanics of Kerimasi, from where it continues south again till Engaruka, where the height of the escarpment is about 550 m. South of Engaruka two fault systems are developed: the main boundary fault of the Manyara basin continuing SSW and a diverging, SE-trending fault, the Engaruka Basin fault, the southwestern boundary of the Engaruka basin. The resulting triangle, the Engaruka Block, is back tilted SSW towards the Manyara boundary fault, broken by minor faulting, and decreases in height south of Engaruka from (1220 m a.s.l.) towards Lake Manyara (945 m a.s.l.). The SE trending fault system results in many faults, creating horsts, graben and fault blocks. This trend is visible over a larger area (examples, Lembolos graben, Oljoro Graben, many of the minor faults north and south of the Arusha-Makuyuni road). This fault system has been responsible for the formation of many smaller volcanic cones.

Fig.1: Escarpment characteristics of the EAR valley in the western side as a part of a continuous half-graben in Northern Tanzania

Figure 2: A scaled photo showing escarpment characteristics of the EAR valley in the western side as a part of a continuous half-graben in Northern Tanzania .

Geological evidences suggest that Natron Basin was connected to the Magadi Basin in Kenya (Dawson, 2008). Its estimated depth is about 3.3 km (Ehbinger and others, 1997), which corresponds with the 3.5 km depth of Lake Magadi in Kenya, obtained from seismic refraction data (Birt and others, 1997). Sedimentation in the proto-lake Natron began around 1.7 – 1.8 Ma, and was probably bounded by faults. These Peninj Group sediments, named after the Peninj River, where they are well exposed at the Natron fault area, lay above basalts and nephelinites of Ol'Doynio Sambu and Mosonik volcanoes, which are dated at 3.5 – 1.77 Ma (Isaac and Curtis, 1974; Manega, 1993). The separation between both is a thin sedimentary layer and the weathered Hajaro lava layer. The whole sequence is called the pre-Peninj Beds (Dawson, 2008). In the Natron fault boundary, forming the half graben (Fig. 1 and 2) where present-day lake Natron is situated, the beds are exposed as a 100m succession of sediments and intercalated lavas.

Recently Sherrod and others (2013) studied old Lake Natron beds at the southern side of the present lake, which show thin-bedded gray sandstone and white siltstone exposure at an altitude of 620-640 m, southeast of the lake. Blocks of carbonate, probably of stromatolitic origin, are also observed. Sand- and siltstones are certainly lacustrine in origin. These are the most southern deposits known for Lake Natron, even though the lake may have extended and connected with the Engaruka Basin (Dawson, 2008) before the onset of Kerimasi and Ol'Doynio Lengai volcanics. Some large debris-avalanche blocks from Ol'Doynio Lengai are interspersed with the lacustrine deposits.These debris ripped into the lacustrine deposits and deformed them. The carbonate layers shattered but the sand- and siltstone layers folded (Sherrod and others, 2013). Dating by Sherrod and others (2013) of these lacustrine beds shows ages between 240.000 and 135.000 years, based on U/Th ages from stromatolites which mark a shoreline altitude of 655 m (Hillaire-Marcel and others, 1986). They also described two different erosional surfaces above these ancient lake Natron Beds, which led them to conclude the following geomorphological sequence of deposits and erosion for this southeastern area:" Lacustrine sedimentation of ancestral Lake Natron continues for some period of time following incursion of debris avalanches into the area, but ultimately the lake recedes and an alluvial pebbly sand is layered across the slope. Erosion isolates this pebbly-armoured surface, perhaps because the basin center to the north subsides more greatly than the southern margin where these outcrops lie. The encroaching Gelai sandstone blankets the area; but then it too erodes back, to form the setting found today".

The present-day lake (Fig.3) is surrounded by, and sits below, the High Natron Beds, a 7 m zeolitized section of clays and tuffs with irregular nodes of chert, some rimmed by magadiite and kenyanite. The Beds can be correlated with the high magadi beds around lake Magadi, deposited during a lake highstand around 9 ka (Butzer and others, 1972). There was an earlier highstand around 135 ka, based on freshwater stromatolite growth (Hillaire-Marcel and others, 1986). The modern lake is 610 m

a.s.l., it is 22-35 km wide and 75 km long. It is shallow (only 3-4 m deep). Evaporation rate is very high (up to 20mm/day). The saline waters are rich in sodium carbonate/bicarbonate, NaCl and silica, and the pH is more than 9.5 (Howell, 1972).

There are 4 permanent rivers feeding the lake:1) Engare Nyiro, 2) Peninj,3)Moinik, 4) Engare Sero, all rising from the western highlands, entering the lake from its western (Engare Nyiro, Peninj, Moinik) and southwestern side(Engare Sero) . Apart from the Engare Sero river which brings only volcanic material from the Crater Highlands, the others also bring metamorphic material from the basement hills west of the Nguruman and Sambu escarpments. Where the rivers enter the lake, there are brackish lagoons, and there are also brackish and salt springs. When the lake reaches high salinity, there is a proliferation of brine shrimps, responsible for the bright-red coloration. A big part of the lake surface is covered by trona, thermonatrite and halite. Below the surface, similar layers are interbedded with black organic clays (up to 6 m deep), resulting from volcanic dust input, which, combined with organic material from detritus and algae, form the organic rich bottom sediments. During eruptions of Oldoinyo Lengai, there was direct volcanic input. Also, magnesite deposits around one of the springs at the east side below Gelai is visible.

Fig. 3: Lake Natron and formal Lake Natron deposits with the associated geomorphological features as observed from the western side of the East Africa Rift System

4. Mount Ol'Doynio Lengai

Ol'Doynio Lengai (2952 m) is a stratovolcano, the only active volcano in Northern Tanzania, characterized by steep slopes. It is the only world's active volcano emitting carbonatite lavas. The diameter at the volcano's base is around 12 km, and the very symmetrical shape of the volcano results from an eruptive activity centered near the summit area. Several tephra cones and tuff rings are found on the lower flanks of the volcano. Some of these are probably related to the magmatic system which built the main volcanic body (Keller and others, 2006). The volcano is mildly incised by radial gullies, with exposures 30 to 50 m deep. The eastern and southeastern flanks are more incised because, due to the prevailing winds, more fallout material covered north and west flanks and the active vent of the volcano moved slightly northwards over the last thousands of years. On its northeastern upper slopes, a major landslide scar, 300 m deep, is visible. This scar is known as the Eastern Chasm (Dawson, 1962; Kervyn and others, 2008a). Debris avalanches and flows from this landslip occur in the plains to the NE and east. Others originating on the northern slopes have created deposits on the plains north of the volcano, where some of them formed low islands in Lake Natron some 16 km to the North. C_{14} dating on plant rests in some of these avalanche deposits have been done by Klaudius and Keller (2006), resulting in an age of 2500 years BP.

Historical development: Before 1960, there was no proper documentation on volcanism at Mt. Ol'Doynio Lengai. The records indicate that from 1960, the depth and shape of the active crater has been changing. The first change is noted during the partial infill by ash in the 1966 eruption. This was followed by partial collapse of this ash cone to form a pit crater. Between 1967 and 1983 there was no activity, but, after the renewal of the activity in 1983, the pit crater was slowly filled by lava flows and spatter from a series of small transient cones, hornitos and lava pools. In 1993, a small ash eruption in which the lapilli were of mixed carbonate-silicate composition accompanied extrusion of exceptionally voluminous and viscous natrocarbonatite lava flows, containing globules of silicate lava (Dawson and others, 1994). In November 199, the crater became almost infilled by lava and spilled over (Nyamweru, 1999). Afterwards, minor lava extrusions and overspill have continued till at least August 2007. A major hornito collapse occurred into an underlying lava pool that caused a lava surge, coming down the western slope on 29 March and 3 April of that year (Kervyn and others, 2008a). Between 1998 and September 2007, a crater wall was present only on the south side below the summit, and the crater was, as in 1904, a carbonatite platform with hornitos and lava pools. As a result, from the September 2007-April 2008 eruptions, the former crater is occupied by an ash cone with a central ash pit.

Fig. 4: Debris from the EAR Escarpment, vegetation at Engaresero and Mt. Ol'Doynio Lengai viewed from the North.

Ol'Doynio Lengai (Fig.4) is a volcano with explosive eruptions. It was formed by tephra fallout, pyroclastic flows, debris avalanches and flows. Lava does not represent more than 10% of the cone (Dawson, 1962). The main type of eruptive material is nephelenite and phonolite, with very peralkaline rocks. Carbonatite lava represents only a tiny part of the volcano, less than 1 % according to Donaldson and others (1987) to less than 5% according to Klaudius and Keller (2006).

Dating records: Sherrod and others (2013) published new radiometric ages for volcanic material from Ol'Doynio Lengai. Four samples were analyzed, two from the volcano itself and two from debris avalanche deposits downslope, by Ar40/Ar39 incremental- heating techniques. Ages of 338± 23 ka from a phonolite lava flow on the eastern lower slope and 353± 65 ka from a tuff ring in the same area reveal that the volcano was growing around 350 ka, with well-established lower slopes, with posterior growth of the steep main cone. The samples, taken from the debris avalanche deposits north of the volcano, reveal astonishing results. Dates of 793± 63 ka from a phlogopite mica-rich tuff layer in an older debris avalanche deposit, and 460± 75 ka also from a phlogopite tuff in a younger debris avalanche deposit, must be interpreted as of an ancestral Ol'Doynio Lengai, completely hidden under the modern volcano. This ancestral volcano must have had a period of building up and failure – maybe more than once. If these dates are related to Ol'Doynio Lengai, which, as the authors claim, is not certain but possible, it would mean the volcano was active for much of the last 500000 to 800000 years. Former dating up to

350000 years are at least “shaken”. More research in the field is definitely necessary, as two datings for such an important feature seem low.

5. Evidences for East Africa drifting

At the border between Tanzania and Kenya on the western side of Lake Natron, there is a shield volcano (Ol’Doynio Sambu) cut by the EAR system indicating that the volcano is older than the rift. The observed hills to the East were formally part of the escarpment, their current position being evidence of subsidence and widening of the rift.

Close to Mt. Ol’Doynio Lengai on the Southern side of Lake Latron(Armykon Hill), two lava flows are visible. They seem to be separrated by an inferred graben later filled by alluvial sediments

Sink pits: There are very recent sink pits close to Mt. Ol’Doynio Lengai(Fig. 5) indicating that the EAR System is currently stretching. According to Local people, such pits were developed between 2008 and 2015.

Fig 5: Sink Pit within the East Africa Rift system with Mt. Ol’Doinyo Lengai view as observed from the N .

6.0 Mlima wa Nyani (Ol'Doynio Loldulal)

This is a hoodoos structure probably modified by differential erosion (Fig.6 and 7). It was in this area the oldest dating in a debris avalanche deposit was recorded for Ol'Doynio Lengai by D. Sherrod

Fig. 6: Mlima wa Nyani

Fig.7: Mlima wa Nyani with view of Ol'Doynio Lengai and Kerimasi volcanoes

7.0 Engaresero Hominid Footprints

In the Engaresero area hominid footprints were discovered dating back 120,000 years ago (Fig. 8 to 10). They were discovered in 2006 by Koongo Ole Sakai who later reported to Julian Von Mutius (the owner of Engaresero Camp). The footprints are the best-preserved track ways of early modern humans. Preservation was probably favored by rapid accumulation of volcanic ashes and alluvial sediments on top of them. Other significance of these footprints include

1. the rarest human/homindi foot prints in Africa
2. The present multiple individual foot prints from adult to juveniles

Fig. 8: Engaresero hominid footprints area. Close to the information board (right) is the Maasai who discovered these footprints

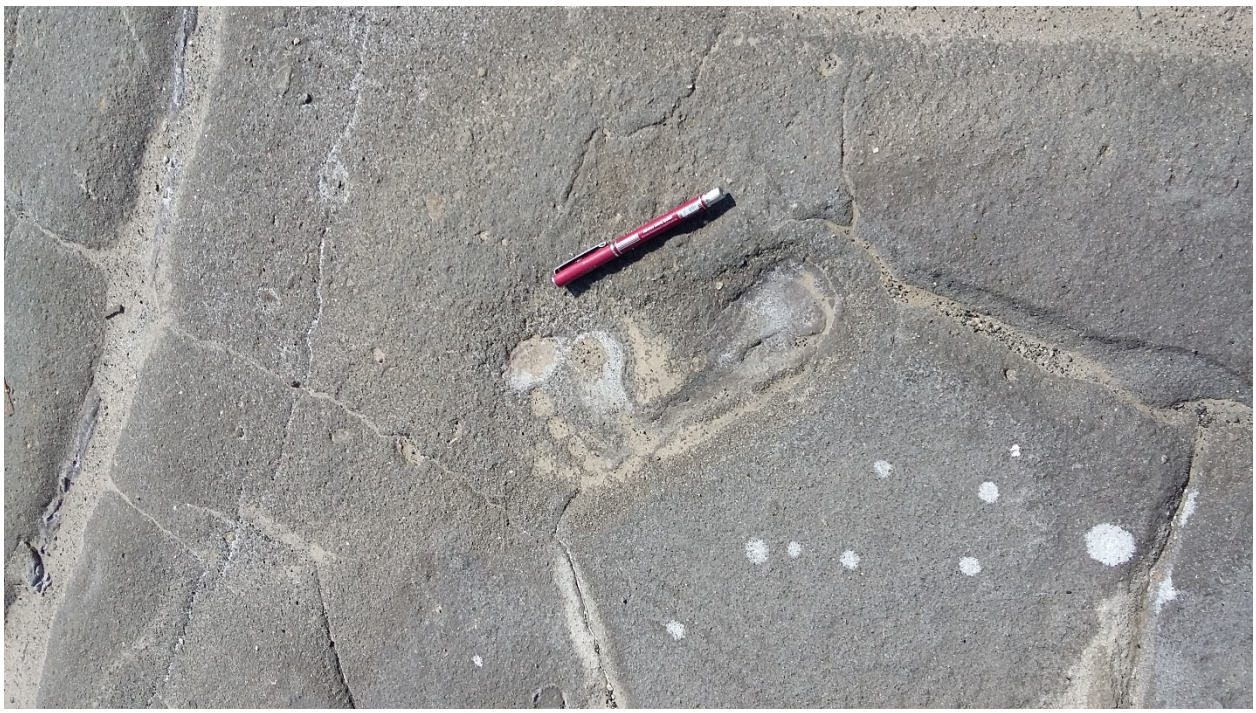

Fig. 9: A scaled Engaresero foot print

Fig. 10: Students observing carefully Engaresero foot prints

8.0 Volcanic Hazards

The following are the common identified volcanic hazards

1. Air fallout or tephra fallout
2. Pyroclastic flows
3. Lahars
4. Tsunamis
5. Earth deformation
6. Lava flows
7. Gases

Air fallout or tephra fallout: Tephra is the collective term applied to all air fall material like ashes, lapilli and blocks or bombs. Explosive volcanic eruptions produce huge environmental noise and cover the surrounding area with volcanic ash. The thickness and the area covered depends on the intensity of the eruption and the meteorological conditions. With heavy winds the smaller ashes can be swept away, only to later appear as acid rain up to hundreds of kilometers from the volcano. Ashes initially form a hot, ascending ash cloud that rises under extreme pressure. Depending on the explosion, the gases and particles can pass the tropopause (10 km altitude) and can even enter the stratosphere (10-50 km altitude) at up to 25 to 50 km. The smaller ash particles will disappear from the stratosphere in a few months, but sulfurous gases, like sulfur dioxide, can dissolve in atmospheric water drops and react with oxygen to form sulfuric acid. These droplets are called aerosols, and they are capable of orbiting the earth for a long time. These aerosols absorb solar radiations and reflect them back into space, and this can cause unpredictable results that are sometimes spectacular. The eruptions of Krakatau (1883), St.Helens (1980) and El Chichón(1982) all resulted in beautiful reddish purple sunsets. However, the 1991 eruption of Pinatubo volcano in the Philippines showed the other aspect of the action of aerosols by drastically lowering temperatures in the surrounding area.

During those explosives eruptions, part of the eruptive tephra falls down on the volcano and on the adjacent area. The largest fragments fall at high speed close to the vent, while the smaller particles are lifted by gases and carried away from the volcano. How far they travel depends on the type of explosive activity and the meteorological conditions.

Basaltic volcanoes are usually less tragic. Basaltic magmas produce lapilli and Pelee's hair (fine glassy elements resembling hair). These seem less dangerous, but when they fall on fields and combine with gases the results can be tragic. An example of this is the 1783 eruption of the Laki volcanic area in Iceland when nearly all the local livestock died after eating the grass which had been contaminated with Pelee's hair and gases. This resulted in a famine that was responsible for the death more than 10,000 people. Fortunately, a case like this would be much less tragic today due to modern communications and better knowledge. On the bright side, these ashes can have also a beneficial effect on food production. In tropical areas, the combination of water and heat transforms the ashes into fertile land; this is the reason why so many people live close to volcanoes.

Pyroclastic flows: These flows are produced by explosive eruptions or the collapse of a growing dome (Unzen, Japan, 1991). Actually, they are a high-density mixture of hot gases, ashes and blocks of all sizes, which move by gravity at very high speeds - sometimes up to several hundreds of kilometers per hour! Nothing can stop them. They wreck everything on their path. They are the most dangerous volcanic phenomena in the immediate vicinity of the volcano. If the people who live in one of these danger areas are not warned in advance, then a catastrophe is inevitable. The most famous and tragic example is the Mt. Pelée eruption in 1902 in Martinique. The entire city of St. Pierre was destroyed and 29000 people lost their lives. By that time, there was no volcano observatory nearby and people had no scientific

knowledge. Yet in the months leading up to the eruption -and even for years before - there had been clear signs of volcanic activity. It is very likely that the severe loss of life could have been prevented if the authorities would have been alerted and dealt with the situation in time.

Lahars: The Indonesian word “lahar” -or debris flow- is used to describe a mixture of water, volcanic ash, rocks, and even parts of the land, which rush -usually- through river beds at speeds of more than 50 km/hour with a considerable destructive force. These can originate in several ways, and can even start when a heavy rain on ash deposits on unstable slopes puts everything in movement. The volcano does not even have to erupt! This was the case at Pinatubo after the June 1991 eruptions happened at the same time that a typhoon devastated the area. Every year thereafter the area experienced several lahars during the monsoon months. In July 1995, 15000 people had to be evacuated.

The expulsion of water from a crater lake can also create a lahar. This was the case in the eruption of the Kelut volcano (Indonesia, 1919). Lahars can also occur when volcanic eruption material fills a nearby lake, thus triggering a flood of debris.

A fusion with ice and snow creates lahars, as was the case of Nevado del Ruiz, Columbia, in 1985. Even though the eruption was fairly unimportant, with a production of only 0.03 km3 of magma, the fusion with the icecap created a lahar, which in no time destroyed the village of Armero, 50 km away from the volcano. 25000 people died. The record in distance of this type of lahar happened in 1877, when Cotopaxi volcano (Ecuador), erupted and material arrived 240 km from the mountain. Another type of lahar is created by sub-glacial eruptions.

Under a heavy icecap, magma is extruded between the ground and the base of the icecap or glacier, usually in the form of a sill. Even though the magma may melt through the ice and result in a hydromagmatic explosive eruption, the real danger for the area is the escaping water from under the icecap. These floods are called “jökulhaups”, common in Iceland. An important example is the 1996 Gjalp fissure eruption under the Vatnajökull icecap in Southern Iceland. Over a 5 week period a 3.5 km3 of water accumulated under the ice and was released from under the ice in just 2 days, creating a jökulhaup which destroyed roads and bridges before reaching the ocean.

Earth movement (Ground deformation): Explosive volcanoes have pronounced unstable slopes, due to hydrothermal alteration in its interior. When magma rises in such a volcano, it will produce deformations which might be responsible for a partial collapse of the volcano. This creates what is called debris avalanches. These avalanches are very destructive and can carry a long way. In the case of the 1980 eruption of Mount St.Helens -Washington State- a deformation of about 100 m at the northern side of the volcano was formed in 2 months time, and after collapsing , the debris reached a distance of about 25 km through an existing valley, destroying everything on its way (more in Voight and others, 1981). It is

considered that the distance these avalanches can normally travel is about equal to ten times the height of the top of the volcano (measured from the surrounding area). The longest known debris avalanche reached a distance of about 40 km (Socompa volcano, Chili). Mount Meru features one the longer debris avalanches, reaching a distance of more than 30 km east of the volcano. Debris avalanches, like lahars, can occur even without eruptions. Heavy rainfall, combined or not with seismicity, can provide the conditions for heavier volcanic material lying on lighter, maybe weathered material, to start moving, triggering off a landslide.

Tsunami: Tsunamis are big waves that flood coastal areas. They can be provoked by seismic activity related to volcanic activity, submarine explosions, pyroclastic flows finishing in the sea, earth movements and collapses and phreatomagmatic eruptions. The most recent examples that we are all familiar with are the 2004 tsunami in the Indian Ocean, and the 2011 one in Japan. The first resulted in the staggering loss of 230,000 lives and devastating loss of property. The second claimed 16000 victims and has caused a very dangerous radioactive disaster.

Lava flows: Fluid basaltic lava flows can reach speeds up to 100 km/hour. The speed depends on several factors such as volume, slope, temperature, chemical composition, and formation of lava tunnels where lava can flow without losing temperature. Depending on the conditions, these lava flows can travel extremely rapidly or only a few meters per hour. Lava flows are not generally dangerous to people, but they destroy all property in their path. In Hawaii, since 1983, the Kilauea volcano has been producing flows which have destroyed roads, houses and plantations before reaching the ocean. The lava flows are actually increasing the size of the Big Island of Hawaii. One of the biggest recent material disasters happened in 2002 in Goma (Democratic Republic of Congo) where lava flows from Nyiragongo volcano destroyed huge parts of the City of Goma.

Volcanic gases: During an eruption, or even between eruptions, gases escape in the atmosphere. Even during a small eruption thousands of tons of SO_2 can be produced, and when combined with water it creates sulfuric acid. For instance, in 1995, the Popocatepetl volcano (Mexico) released between 2,000 and 3,500 tons of gases every day! Before the 1991 eruption of Pinatubo, during the month of May, its fumaroles produced 500 tons per day in the beginning of the month, rising to 5,000 tons per day at the end of the month. It receded to 1,500 tons just before the eruption, and after the eruption, in the beginning of the month of July, a further 1,000 - 1,500 tons daily were released, receding to 20 tons in October. During a cataclysmic eruption between 50 and 100 million tons of SO_2 can be projected in the atmosphere. Normally these gases are diluted fairly rapidly in the atmosphere and are rarely dangerous, but when driven by winds or coming down as acid rain the consequences can be dramatic. The gases contain fluoride and chloride which can contaminate fields and wipe out cattle, as happened in the aforementioned Laki eruption of 1783.

Emissions of CO_2 can be also be catastrophic. This is a very common gas which is heavier than air – about 50% heavier after it has cooled to ambient temperature. Being this heavy means that in calm conditions the CO_2 will descend to the ground or accumulate in craters or cracks where it will displace oxygen. It is odorless so people may breathe it without being aware and this can cause death. CO_2 can escape slowly from shallow magma bodies into crater lakes making the body of water at the bottom denser so the gas stays in the lower depth of the lake. Heavy rain or a landslide can make this stratification unstable, and the water in the lake can rise up. The deeper water, saturated with gas, will rise to the upper part of the lake where the pressure is lower. The water then becomes oversaturated with the gas. Since this gas is heavier than air it will spread and travel downhill. An example is Lake Nyos, (Cameroon ,1984), where animals and people around the lake died of asphyxiation.

A potentially very dangerous area is Lake Kivu, a large rift valley lake, between Rwanda and the Democratic Republic of Congo. Around 200 km^3 of CO_2 and methane are believed to be contained in its waters. This could lead to explosions and eventually result in tsunamis.

Water containing fluoride is a significant problem in Northern Tanzania where a city like Arusha is situated at the foot of Mount Meru, and where the fluoride content of the groundwater is extremely high. The World Health Organization recommends a permitted maximum of 1.5 mg of fluoride per liter of water, whereas the Tanzanian Government tolerates a much higher percentage of 4mg/l (which at least is lower than their former standard of 8 mg/l). A lot of scientific work has been carried out to measure the fluoride content in the Mount Meru area, where fluoride readings in mg/l up to hundreds have been recorded (e.g..Ghiglieri and others, 2010).

9.0.Volcanic hazard monitoring

The monitoring of a volcano is based on geophysical and geochemical studies of the phenomena which accompany its eruptive mechanism. It consists of equipping it with specialized devices that are distributed on the volcano in such as to achieve the best results for the different parameters. Monitoring aims to answer the questions of where, when, how, which type of eruption? What is the danger for the population? To answer these questions, we need specific studies of the contemporary eruptions, reconstruction of former eruptions, modeling and theoretical simulation of eruptions, etc… From these studies, we can make a thorough evaluation of the volcanic risks from which emergency plans can be formulated. In terms of civil protection this is called global previsions or longterm previsions. The following are the general monitoring techniques:

Nets of seismic monitoring: Most eruptions are preceded by seismic movement. Only less than 5% of volcanoes erupt without seismicity. On the other hand, seismic movement does not always mean there will be an eruption. A seismic net is formed by placing seismometers in strategic areas. Important factors in placement are the geographical situation and the type of land. For instance, it is better in hard

rock than in ash deposits. Also important is the site from where signals can be transmitted. The more stations there are, the better the approximation. For instance, at the Piton de la Fournaise, a shield volcano on the island of La Réunion, there are about 40 seismic stations and, according to personal information, this number will soon increase considerably . In Tanzania, potentially dangerous volcanoes like Mounts Meru and Kilimanjaro have no stations at all.

Nets of land deformation: Pressure in the magma chamber and the movement of magna can provoke a deformation of a volcano, and after an eruption can have the opposite effect like the formation of calderas. Deformation results in a higher inclination of slopes, formation of fissures, growing craters, and in the case of acidic volcanoes, huge “swellings”. After an eruption, it is more difficult to see a volcano react and apart from caldera formation, the effects have to be measured with specialized equipment. Monitoring nets of land deformation provide data on the variations of the geometry of a volcano using several sophisticated devices. These includes:

A. **Clinometric nets:** A clinometric net is formed by a series of stations. Each station has 2 clinometers or tiltmeters - one in a radial and the other in a tangential (perpendicular to the radial one) position. This combination of positions allows to determine the orientation of the movement of the side of the volcano in that spot, as well as the intensity of this movement. The volcanologist can then reconstruct what is called “vector deformation”. Tiltmeters give measures in microradians. A microradian measures the variation of a slope which corresponds with 1 mm in height over a distance of 1 km. All this is automatically transmitted to the observatory.

B: Extensometers: Fissures opened or closed are interesting indicators of deformations due magmatic rise. The most sensitive type of extensometer measures up to a thousand million parts of a meter and can only be used on small fissures, a couple of dozens of centimeters wide. For the bigger fissures which are sometimes dozens of meters wide, a stiff measuring tape which cannot be deformed, is connected between two fixed points on both sides of the fissure. The measuring element is part of the system, but is operated manually which can be quite a problem during eruptions! It can measure up to a nanometer.

C: Automatic telemeters: If a volcano is growing, it is logical that the distance between any point of the volcano and a fixed point situated on the exterior will change. An electronic optical telemeter uses laser beams to measure these changes with pinpoint accuracy – measuring distances of several kilometers with a precision of 1 mm per km. The telemeter is connected to a theodolite, and with the appropriate computer program one can measure the distance between the telemeter and several reflectors situated in different positions on the volcano. The coordinates of each reflector are introduced into the program and the laser beam passes through the lens of the theodolite searching systematically for each reflector. Once the laser has detected a reflector the beam returns and measurements are registered. This data is

transferred automatically to the observatory. Most of the volcanoes usually have only 1 telemeter, and their usefulness can be compromised by poor visibility as clouds are a barrier to the laser beam.

D: **Satellite Positioning**: The development of the GPS has been of great help in determining land deformation with an accuracy measured in mere millimeters. The data is sent to the observatory and compared with that of the telemeters.

Gravimetric measurements: Gravimeters measure differences in gravity. If a point is fixed and the underlying land does not suffer modifications, the gravity measured will be constant. When magma migrates in a volcano there is a transfer of material. These materials, which are exposed to high temperatures, have a different density than the surrounding rocks, and thus the gravity will be different. A gravitational station consists of a gravimeter and a GPS receiver. If the altitude of a point varies, the gravity will also change, but without the GPS it would be impossible to ascribe this change in gravity to the transfer of material. Automatic transmission of data to an observatory is possible.

Nets of magnetic monitoring: It is normal for the earth's magnetic field to change over time. These variations are very heterogeneous on the surface of a volcano, due to the complex internal structure of a volcano. However, deep inside the volcano the magma creates high pressure with resulting fissures in the areas where the gases circulate. This brings with it a series of processes that are magnetic in nature, and the magnetization of the rocks can be changed due to these tensions. This creates what is called induced magnetic fields. The monitoring station measures the result of all these fields - including the earth's magnetic field. One reference station is located in an area not influenced by the magnetic field of the volcano. A station is comprised of a magnetometer combined with an electronic device which sends the results to the observatory. It is a very complex system, but it is the only way to measure the magnetic variations induced by eruptions. Not many volcanoes in the world are equipped with such a sophisticated magnetic net.

Nets of radon monitoring: All volcanic rock contains U_{238}. The quantity is usually about one to two grams per ton for basaltic and andesites, more for differentiated lavas. Due to radioactive disintegration the uranium gives off a series of radioactive elements like radium and radon. Radon has a half-life of 3.65 days, meaning every 3.65 days half of the radon disappears by disintegration if separated from its "mother element", in this case radium. This means that if magma takes 2 weeks to surface then the radon in it can no longer be measured. So why do we use radon for monitoring? There are several reasons: radon is a radioactive gas, easy to measure; it is present in the rocks; water vapor and CO_2 escaping from the magma chamber bring it to the surface. The measuring probe is put into a plastic pipe at a depth of about one meter. The site for the probe must be well chosen in an area where the diffusion

of radon and CO_2 are at their maximal level. The infrastructure for it (batteries & solar panels) is the same as for the geophysical stations, and the data is sent automatically to the observatory. These stations are relatively inexpensive to create, so it is practical to put as many as possible on a volcano to obtain the most accurate information about an active episode. The “radon ascent" is not uniform in a volcano, but is related to the localization of the injection of magma. This means one has to be in the right place at the right time to get useful results.

Nets of hydroacoustic monitoring: Vibrations are produced in a volcano by the circulation of lavas and fluids, plus the micro-seismicity due to micro fissures which precede the great fractures in the area of the magma chambers. The attenuations of these vibrations are less profound in a liquid environment than in the heterogeneous volcanic rocks. Therefore, one can put a hydrophone in a crater lake, and this will provide a good connection between the device and the volcano. Frequencies from as low as 0.1 Hz up to ultrasonic frequencies (200.000 Hz) can be measured. Volcanoes like Kelut (Indonesia) and Taal (Philipines) are equipped with this type of monitoring. It is a promising application for the future in sites where there are crater lakes or perforations in appropriate sites.

Nets of geochemical monitoring: Magma is a molten mass of silicates mixed with dissolved gasses and minerals. With enough pressure, fissures will open and gasses can escape through fumaroles and fissures, before the arrival of magma. The oldest system used was to measure the temperature of the gases in fumaroles or emitted from phreatic layers, heated by magma. A thermometric probe is installed in fumaroles and fissures, and an electronic device sends the data to the observatory. This seems easy, but sometimes these devices have to be put into hostile areas, and they measure high temperatures of pure acid of several hundreds of degrees. Corrosion problems are endemic, and the probes break down quite often. As far as the gases themselves are concerned, the common to monitor are carbon dioxide and sulfur dioxide. Several methods are used. Gas samples can be taken in small containers from the soil or directly from fumaroles or fissures. They are then brought back to the laboratory for analysis. This system can be tricky as the gases are hot and sometimes the sampling sites are hazardous. After the cooling of these gases there is an additional risk of a chemical reaction between the different gases if several are present. It is much easier to measure these gases using spectrometers either on the ground or flown in airplanes above the volcano. A spectrometer measures the amounts of CO_2 and SO_2 in the atmosphere by detecting the amount of light absorbed by these gases. Correlation Spectrometers (COSPECs) work at ultraviolet wavelengths. Fourier Transform Infrared Spectrometers (FTIRs) work in the infrared wavelengths.

Nets of manual measuring: Some measurements are still done manually and there are two primary types of manual nets: geophysical and geochemical. Geophysical nets are created in the field by using fixed lead or steel points (usually 4 around a central one). For land deformation, theodolites are

used as in normal topography. From the central point the heights of the other points are measured looking for differences. The same points can be used for magnetic and gravity measurements. Geochemistry involves regular measurements of chemical composition. The rising of magma and the fissures that are formed by its pressure in the higher part of the magmatic chamber causes a differential degasification - a variation in the chemical composition in the fumaroles. Volatile gasses like CO_2 and SO_2 get through the phreatic layers and contaminate. Fumaroles are monitored regularly, at least once a month – and even more often during active periods. It is important to understand that the major component of the volatile elements is water vapor (often more than 90%). Only a small part of this water comes from the magma itself as most of it is infiltrated water mobilized by contact with magma. Temperature variations in fumaroles are significant factors. The closer magmatic gases get to the surface, the more the temperature in fumaroles will rise. So, it is important to have regular readings of these temperatures.

Conclusions and Recommendations

The field excursion to northern Tanzania, along the East Africa Rift System has enhanced the volcanic knowledge to students, particularly in tectonics, volcanic style, types and associated geomorphologic features, potential natural hazards and monitoring techniques. Very special interest to students were the Engaresero foot prints (dated 120,000 years) and current geological features which indicate the stretching of the East Africa Rift Valley (sink pits and hills/cones cut by the rift).

Based on both lectures during the excursion and field trips, the study recommends the following:

Creation of awareness on volcanic hazards to local people living in volcanic risk zones. This should be accompanied by the setting of seismic nets for prediction of future events

Systematic dating of materials along the rift for magma evolution and proper reconstruction of volcanic activities along the rift

As the rift system is characterized by different geological formations, more research on material science, energy and minerals are recommended.

Finally, as volcanology courses have been offered at NM-AIST to students with different background/disciplines, it is recommended that the students (particularly those lacking geological background) get an overview or Introduction to geology before field work/ excursion starts.

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