Tectonics of Saturn's Moon Titan AND Tsunami Modeling of the 1629 Mega-thrust Earthquake in Eastern Indonesia

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The Tectonics of Saturn’s Moon Titan

AND

Tsunami Modeling of the 1629 Mega-Thrust Earthquake in Eastern Indonesia

Zac Yung-Chun Liu

A thesis submitted to the faculty of
Brigham Young University
in partial fulfillment of the requirements for the degree of

Master of Science

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ABSTRACT

Chapter1-2: The Tectonics of Saturn’s Moon Titan

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The Cassini RADAR mapper has imaged elevated blocks and mountains on Titan we term ‘ridges’. Two unresolved problems regarding Titan’s surface are still debated: what is the origin of its ridges and was there tectonic activity on Titan? To understand the processes that produced the ridges, in this study, (1) we analyze the distribution and orientation of ridges through systematic geomorphologic mapping and (2) we compare the location of the ridges to a new global topographic map to explore the correlation between elevation and ridges and the implications for Titan’s surface evolution.

Globally, the orientation of ridges is nearly E-W and the ridges are more common near the equator than at the poles, which suggests a tectonic origin for most of the ridges on Titan. In addition, the ridges are found to preferentially lie at higher-than-average elevations near the equator. We conclude the most reasonable formation scenario for Titan’s ridges is that contractional tectonism built the ridges and thickened the icy lithosphere, causing regional uplift. The combination of global and regional tectonic events, likely contractional in nature, plus enhanced fluvial erosion and sedimentation near the poles, would have contributed to shaping Titan’s tectonic landforms and surface morphology to what we see today.

However, contractional structures (i.e. thrusts and folds) require large stresses (8~10 MPa), the sources of which probably do not exist on Titan. Liquid hydrocarbons in Titan’s near subsurface must play a role similar to that of water on Earth and lead to fluid overpressures, which enable contractional deformation at smaller stresses (<1MPa) by significantly reducing the shear strength of materials. We show that crustal conditions with enhanced pore fluid pressures on Titan favor the formation of thrust faults and related folds, in a contractional stress field. The production of folds, as on Earth, is facilitated by the presence of crustal liquids to weaken the crust. These hydrocarbon fluids have played a key role in Titan’s tectonic evolutionary history, leaving it the only icy body on which strong evidence for contractional tectonism exists.

Keywords: Titan, Saturn, Icy satellites, Tectonics, Radar Observations, Structural Geology, Geological Mapping, Fluid pore pressure, lithospheric strength
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Chapter 1:

The Tectonics of Titan: Global Structural Mapping from Cassini Radar

1. Introduction

The Cassini spacecraft’s 2.17 cm RADAR instrument has revealed remarkably Earth-like processes that have shaped the surface of Saturn’s largest moon Titan (Elachi et al., 2005; Lopes et al., 2010a), such as aeolian (Lorenz et al., 2006; Radebaugh et al., 2008; Radebaugh et al., 2010; Savage et al., 2014), fluvial (Lorenz et al., 2008; Burr et al., 2009; Langhans et al., 2012; Burr et al., 2013), lacustrine (Stofan et al., 2007; Hayes et al., 2008; Lorenz et al., 2014), cryovolcanic (Lopes et al., 2007; Lopes et al., 2013) and tectonic (Radebaugh et al., 2007; Radebaugh et al., 2011; Liu et al., 2012; Solomonidou et al., 2013; Liu et al., 2014). These processes have formed and shaped ubiquitous Earth-like surface features and landforms on Titan. The features are RADAR bright as seen by Cassini’s Synthetic Aperture Radar (SAR) and have relatively high elevations we refer to as mountains (Radebaugh et al., 2007; Barnes et al., 2010; Liu et al., 2012) and hummocky terrains (Lopes et al., 2010a), which have been interpreted to be related to tectonic processes (Radebaugh et al., 2007; Liu et al., 2012; Solomonidou et al., 2013). The linear/curvilinear surface features have been the indicative of tectonism. On Titan, the degradation plays the important role to alter the surface landscapes. Thus, if tectonic stresses have contributed to the deformation of Titan’s icy crust, interpreting the morphology of these tectonic features should be treated cautiously. The possible tectonic landforms, such as mountains, ridges, and faults, can be examined in geomorphological and structural mapping.
through analysis of high-resolution Cassini SAR images, obtained globally beginning in 2004 (Elachi et al., 2005).

Since Titan’s mountains have a variety of morphologies, in this study, we classified mountains and hummocky terrains into three categories using the terms: (1) *ridges*: chains of hills with elongate, curvilinear/linear morphology and higher than the surrounding areas (Figure 1a), (2) *isolated blocks*: elevated blocks which are generally isolated (Figure 1b), (3) *rugged terrains*: rugged mountain chains which possibly have been through erosive processes and have a hummocky morphology, wherein multiply adjacent peaks extend across vast regions (Figure 1c). Rugged terrains are mainly located in the Xanadu region (centered at 5° S 100° W), the first surface feature of Titan seen from Earth (Lemmon et al., 1993; Smith et al., 1996). Xanadu stands out globally as a bright feature on Titan’s leading hemisphere and this brightness is the result of either compositional or textural differences in this region compared with other areas on Titan (Radebaugh et al., 2011; Langhans et al., 2013). On Titan, the longest ridge belts (> 100 km long) are mainly located in the equatorial regions (Liu et al., 2012). Impact craters are also SAR bright (Figure 1d) but have a highly curved morphology, in contrast with the curvilinear morphology of ridge belts. Based on the difference in morphology, one can distinguish impact craters from mountainous terrains on Titan. There are a few impact craters, though far fewer than could be expected compared to other bodies in the solar system (Radebaugh et al., 2007; Wood et al., 2010). The lack of craters, destroyed by resurfacing inclusive of erosion, indicates that Titan’s surface is very young; between 100 and 500 million years old (Lorenz et al., 2007); Wood et al, 2010). In addition, Cassini’s Imaging Science Subsystem (ISS) has showed evidence that seasonal precipitation (e.g. methane rainfall) has facilitated the erosional process on Titan’s surface (Turtle et al., 2011a; Turtle et al., 2011b).
Currently, two unresolved problems regarding Titan’s surface are still in debate: what is the origin of its ridges and was there tectonic activity on Titan? To be able to generate the tectonic features seen on the surface, sufficient internal energy is required, which is related to the degree of the differentiation of Titan’s interior. The relatively low moment of inertia (MoI = 0.34) observed by the Cassini spacecraft (Iess et al., 2010) indicates that Titan’s interior is only partially differentiated and therefore the geological activity produced by internal heat must be very limited (Nimmo and Bills 2010; Hemingway et al., 2013). However, O’Rouke and Stevenson (2014) found that convection cannot realistically remove radiogenic heating over geological time, so Titan must be differentiated. They state that the discrepancy in the moment of inertia could be explained by Titan having a mantle of serpentinized rock. For icy satellites, tidal forces and rotational mechanisms (e.g. disspinning, polar wander) may produce stresses that can cause crustal deformation and these stress mechanisms play important roles in forming tectonic patterns (Collins et al., 2009). Thus, analyzing Titan’s tectonic patterns would help us to understand its geological history and is one of the main purposes of this study.

Two broad hypotheses have been extended for the processes that produced Titan’s ridges: exogenic and endogenic. Moore and Pappalardo (2010) proposed the exogenic hypothesis, pointing out that the ridges on Titan are similar in scale to those on Callisto, which owe their origin to multi-ring basin-forming impacts followed by erosion. However, the linear-to-arcuate ridges with relatively higher elevation on Titan’s surface (Paganelli et al., 2010; Radebaugh et al., 2011; Cook et al., 2012; Liu et al., 2012) have morphologies consistent with extensional or contractional tectonism (Radebaugh et al., 2011; Figure 1a), which implies an endogenic origin.

Analyzing topographic data and undertaking global mapping of surface features are the keys to testing a possible tectonic contribution to shaping Titan’s surface (Moore and Pappalardo
2010). However, no previous studies have focused on systematic geomorphologic mapping and quantitative analysis of structures and their orientation and distribution on a global scale on Titan. In addition, the driving forces of tectonism and the tectonic evolutionary history of Titan still remain unclear. Thus, the purpose of this study is to test the hypothesis of the origin of ridges by: (1) analyzing the distribution and orientation of mountain ridges to reveal their global tectonic pattern; using global structural mapping as a way of determining degree of linearity of surface features and if linear features show evidence of crustal thickening or thinning, and (2) comparing the location of the ridge belts to a new global topographic map (Lorenz et al., 2013) to explore the correlations between elevation and ridges and the implications for Titan’s surface evolution history.

In this paper, we first introduce the available topographic data extracted from Cassini SAR images, which are used in this study, to provide the first-order ridge height and surface slope estimations. Then, we show our global structural mapping procedure and results of ridge distribution and orientation. Finally, the ridge elevation distribution will be evaluated and tested statistically.

2. Topographic data

2.1 Cassini SAR data sets

A major topographic data source on Titan is Cassini SAR imagery. SAR images are obtained in narrow swaths a couple hundred km wide and related to the flyby trajectory with the resolution of 350 m/ pixel, and to date SAR images cover approximately 50% of Titan’s surface. In this study, the ridge elevation, peak height and average surface slope are obtained by radarclinometry (Radebaugh et al., 2007) and SARTopo (Stiles et al., 2009) techniques, which
will be discussed in the following sections. A recently published global topographic map, which utilizes data from SARTopo, and RADAR altimetry (Lorenz et al., 2013) is used to examine the correlation between ridge locations and their corresponding elevations in this study.

2.2 Radarclinometry

Radarclinometry is an approach to extract topographic information from a single Cassini SAR image; it is dependent on the backscatter expected from the given surface and is based on interpreting brightness and shading variations as slopes, followed by the integration of slopes to obtain topography, and thus can achieve single-pixel resolution. For this technique to yield a quantitatively accurate estimate of topography, the backscatter model used must closely match the scattering behavior of the actual surface (Kirk et al., 2005; Kirk et al., 2006).

We assume the backscatter law for mountainous materials to be $\sigma_0(i) \propto \cos(\text{incidence angle})$, which gives constant backscatter across a range of incidence angles and agrees well with scatterometry results (Radebaugh et al., 2007). Measurements are obtained across mountain peaks of modest widths, typically < 30 km (Kirk et al., 2005). Based on the results from the radarclinometry study of Radebaugh et al. (2007) and Liu et al. (2012), the measured 300 mountains reveal that peak heights range from 120 m to 3300 m, and over 250 peaks exceed 1 km in height.

2.3 SARTopo

In Cassini SAR imaging, multiple antenna beams are employed. SARTopo is an approach to use an amplitude monopulse technique to compare overlapping regions between the beams and thereby estimate surface elevations (Stiles et al., 2009). This technique depends upon the spacecraft attitude and the antenna gain pattern. It enables two or more surface-elevation profiles
to be obtained along the long dimension of each SAR image, each with approximately 10 km spatial resolution and 75 m height accuracy.

SARTopo profiles yield absolute elevation changes of the surface; in contrast, the topographic profile obtained by radarclinometry is the relative height of the ridge ‘peak’ compared with the topography of the surrounding terrain. To be able to estimate the average surface slope for ridges, one should extract the absolute elevations from SARTopo profiles. Based on the results of Liu et al. (2013), the estimated surface slopes of 10 ridge belts, from base to peak, are well below 2°. Compared to the slopes of ridges on terrestrial planets, the surface slopes of Titan’s ridges (<2°) are fairly gentle, which is a strong indicator of the nature of the ridges’ tectonic style (e.g. contractional or extensional) (discussed in Section 3.5).

2.4 Global Topographic Map

A new global topographic map has been produced by Lorenz et al. (2013), in which SARTopo and altimetry data are used to construct a global gridded 1 x 1° elevation map (Fig. 3 in Lorenz et al., 2013). Titan topographic data are sparse, and most of the map domain (90%) is populated with interpolated values using a spline algorithm. The highest (520 m) gridded point observed is at 48°S, 12°W. The lowest point observed (1700 m below a 2575 km sphere) is at 59°S, 317°W. Interestingly, the Xanadu region (centered at 5°S 100°W) is generally lower than the surroundings. In this study, we incorporate this global topographic map with our structural map to examine the correlation between elevation and ridge distribution (discussed in Section 5).

3. Global Structural Mapping

3.1 Mapping method
The geological mapping of tectonic structures can be used to interpret the types or sources of stress related to their formation (Tanaka et al., 2010). The method of structural mapping enables us to determine Titan ridge origins by revealing global tectonic pattern and key morphologies. On Titan, the observed elevated ridges are most likely the results of crustal movement due to tectonic process as well as structures associated with fluvial networks and aeolian erosion (Solomonidou et al., 2013). Black et al. (2012) undertook geological mapping of fluvial features on Titan and their quantitative analysis shows spatially averaged fluvial erosion of 0.4% to 9% of the initial topographic relief and suggests that fluvial erosion is not capable of eliminating mountain ridges. Ridges on Titan have likely still preserved their original trend and orientation to some degree. Additionally, mapped curvilinear surface features and lineaments presumably reflect subsurface structure (Thomas 1988). Thus, regardless of the erosional processes on Titan, mapping observed trends and morphologies of ridges is an acceptable way to examine their structural pattern.

In this study, we mapped SAR-bright ridges as tectonic units and traced the strike of the ridges (Figure 2a) as lineaments across Titan. The resolution of Cassini SAR images (350 m/pixel) is sufficient to map observed curvilinear ridges and lineaments. The SAR images we use here for structural mapping are from Titan flybys Ta-T84, those obtained from October 2004 to June 2012. We mapped the ridges in the equatorial and mid-latitude regions (60° N-60° S) using the USGS Titan ArcGIS project with a geographic projection. For the surface features in the polar regions (60° N-90° N; 60° S-90° S), due to the increasingly distorted images on the geographic projection, we map ridges on polar projection basemap instead. All features are identified in the original SAR images and then mapped on ArcGIS. Each mapped unit is a shapefile of a polyline in ArcMap and contains direction (azimuth) and length. Each unit is
mapped by COGO tools in ESRI ArcMap suite, which serve to account for the spherical effects caused by the change from polar projection to geographic projection in ArcGIS. The length and azimuth data of mapped units are used to make rose diagrams to evaluate ridge orientation and distribution.

3.2 Mapping with SAR images

The SAR images used to characterize Titan’s surface morphologies have brightness variations that represent normalized microwave energy backscattered from the surface. SAR brightness is a function of surface slope, dielectric properties, roughness, and the amount of volume scattering (Stofan et al., 2011). Since there are significant differences between microwave and more familiar optical wavelengths of imaging, one should be cautious in using SAR images to map and interpret surface features.

The SAR images are also obtained using a side-looking geometry. The angle at which the SAR images the target as measured from the horizontal at the antenna is called the depression angle, the complement of the look angle, $\theta_l$ (Figure 3). At the target, local undulations combined with the look angle create the local incidence angle, $\theta_i$. Because of planetary curvature, the look angle does not equal the local incidence angle (Figure 3). Based on Ford et al. (1993), the local incidence angle is:

$$\theta_i = \sin^{-1} \left( \frac{r + H}{r} \sin \theta_l \right)$$

where $r$ is the radius of Titan ($r = 2576$ km) and $H$ is the altitude of the Cassini spacecraft. Normally, based on the Cassini Titan flyby data, $H$ is about 1000 km (Stiles et al., 2009) and the incidence angle is between 10° to 35° (Kirk et al., 2006), which makes the look angle relatively
small ($< 45^\circ$) (Ford et al., 1993) typically ranging from $7^\circ$ ~ $26^\circ$. Imaging radars with small look angles, such as Cassini SAR, enhance the topography at the expense of surface roughness information. In our structural mapping, we map the location of mountain ridges more on topography than roughness. Thus, Cassini SAR images are a good tool for our methods. The orientation of linear features relative to the SAR look direction or azimuth also controls the visibility of the features (Ford et al., 1993) and may affect the orientation of mapped units. Where the illumination is parallel to the features, there is no enhancement of the features. Where the illumination is normal to the features, topographic variations stand out. Before mapping the surface features, we compare available multiple SAR images obtained in the same location to decrease the effect of illumination direction on our interpretation of ridge orientations.

Another factor in SAR imaging, dielectric constant, is a measure of how well electromagnetic waves couple with a material (Ford et al., 1993). Since our mapping focuses on linearity of surface structures, morphology and tectonic patterns, the effect of dielectric constant variations is the secondary factor and can be ignored in our mapping.

### 3.3 Ridge morphologic types

We identified seven different types of ridges based on the degree of symmetry and segmentation (Figure 2a, 2b). For ridges with a scarp-like morphology, we map the strike of the basal scarp as an asymmetrical unit, using the line and teeth symbol. Otherwise, we map ridge crest as a symmetrical unit, using the line without the teeth symbol. For homogeneous ridges longer than 100 km, we consider them ‘long’ features (red) (Figure 2c). For ridges generally shorter than 100 km, we consider them ‘segmented’ features (purple) (Figure 2c). For rugged and hummocky ridges, we consider them ‘eroded’ features (yellow) (Figure 2d). There are a few
possible linear valleys or rifts, which appear SAR-black, likely due to shadowing; we consider them as possible extension features (green) (Figure 2d). The seven ridge groups in our mapping are described in the following sections.

3.3.1 Asymmetric long ridges (Class 1) and symmetric long ridges (Class 2)

In SAR images (Figure 2c), ridge belts are elevated, SAR-bright features with curvilinear margins. SAR-black linear features are generally sand dunes, which curve around obstacles, stop at the margins of higher elevation features and merge around the ridges. A set of large, homogeneous ridge belts longer than 100 km were identified (207° W, 10° S) west of the Huygens landing site (Figure 2c). The heights of these ridge belts are few hundred to 1000 m based on SARTopo profiles (Stiles et al., 2009), with a few slightly higher peak elevations based on radarclinometry (Radebaugh et al. 2007). These ridges found in Cassini flybys T8 and T61 have distinct fault-and-thrust morphologies. Their curvilinear morphology is comparable to the Yakima fold belts in Washington, US (Discussed in Section 5). These ridges are seen to have a slightly asymmetric morphology, in which the ridge has an overall steeper slope on one side than the other (Figure 2c). These are termed ‘asymmetric long’ features and are mapped using the red line and teeth symbol (class 1). Each line represents a single ridge scarp longer than 100 km, and the teeth point in the possible dip direction of fault plane. Similar in general morphology to the asymmetric long ridges, but with generally similar slopes on either side of the ridge, are the ‘symmetric long’ ridges. These are mapped using the red line symbol only (class 2). Each line represents a single ridge crest longer than 100 km (Figure 2c).

3.3.2 Asymmetric segmented ridges (Class 3) and symmetric segmented ridges (Class 4)
Multiple segmented ridges were identified north of the Huygens landing site (202° W, 2° S) (Figure 2c). Based on the homogenous and continuous trend and morphology, these segmented ridges likely used to connect to each other as longer ridges but they were disconnected by fluvial erosional processes and were buried by eolian processes, resulting in the current, segmented components. Due to their geomorphological difference from class 1 and class 2 ridges, here we map the multiple segmented ridge scarps as ‘asymmetric segments’, using the purple line with teeth symbol (class 3). For the segmented ridges have crest morphologies with equal slopes on either side of the ridge, we map them as ‘symmetric segments’, using the purple line without teeth symbol (class 4). Each mapped scarp and crest unit is shorter than 100 km. Compared with the class 1 and class 2 longer scale ridges, the class 3 and class 4 segmented ridges are more common and are distributed all across Titan, while the class 1 and class 2 ridges can only found in the equatorial regions. In addition, based on the radarclinometry measurements (Radebaugh et al., 2007 and Liu et al., 2012), the heights of segmented ridges are generally lower than that of continuous ridges.

3.3.3 Asymmetric erosional ridges (Class 5) and symmetric erosional ridges (Class 6)

The Xanadu region (centered at 5° S 100° W) stands out globally as a bright feature on Titan’s leading hemisphere due to the textural difference caused by the rugged and heavily fluvially eroded morphologies. Xanadu contains evolved river channels, impact craters, and dry basins filled with smooth, SAR-dark material, perhaps sediments from past lake beds (Radebaugh et al. 2011) (Figure 2d). Since Xanadu’s impact crater density is the highest on Titan, it is believed to be the oldest province on Titan (Radebaugh et al., 2011; Woods et al., 2010). Mountains exist in Xanadu as multiple, adjacent, mountain peaks scattered across the terrain in rugged terrain morphologies (Radebaugh et al. 2011). Some mountains can be found in
linear or arcuate, ridge-like patterns. Ridge relief within Xanadu can reach over 2000 m, yet the overall elevation of Xanadu is low compared to surrounding sand seas (Radebaugh et al., 2011; Liu et al., 2012). Even with the complexity of the morphologic features in Xanadu, the trend of arcuate and aligned mountain ridges are still recognizable under Cassini SAR resolution. In Figure 2d, multiple rugged and arcuate ridges with scarp morphologies are mapped as ‘asymmetric erosion’, using the yellow line with teeth symbol. Rugged ridges with symmetric crests are mapped as ‘symmetric erosion’, using the yellow line without teeth symbol. Other than the Xanadu province, the rugged ridges can be found at (75° W 50° N).

3.3.4 Possible Extensional features (Class 7)

There are a few possible linear rift valleys in between elevated ridges, though perhaps due to the intensive weathering and surface degradation processes on Titan’s surface, these are rare. Sand dunes and fluvial sediments have possibly filled in other rift valleys. These rift-like features are typically 150 km long and are mainly located in Xanadu region (Figure 2d) (100° W 20° S) and (240° W 20° S) (Radebaugh et al., 2011). Since these SAR-dark features have linear morphologies different from dunes and are located in between elevated ridges, we map these features as ‘possible extension’, using the green line symbol in our mapping.

3.4 Results

The global structural map is presented in Figure 4. The mapped units are limited to the locations where there is SAR coverage, so the ridges out of swath boundaries are not mapped. A total length of 85,388 km of ridge features were identified and mapped. The pattern of E-W orientation is obvious globally, both at equatorial and polar regions. The long (> 100 km) ridges (red) are mainly located in the equatorial regions and the rugged/ fluvial eroded ridges (yellow)
are mainly located in the Xanadu province. Since the mapped units contain information on direction (azimuth) and length, we use these data to analyze the distribution of the structures and to examine the tectonic pattern in rose diagrams.

3.4.1 Ridge distribution

To analyze the distribution of the mapped ridges, we set up a numeric indicator, the structure density (SD) = 100* Total mapped ridge structure length/SAR coverage area (unit: 100 km/1 km²). Based on the definition of the structure density, a high SD means there are more mapped structures or ridges per unit area. We divide Titan’s surface into six 30° latitude bands:


Based on Table 1, the highest SD is in the south equatorial (SE) region between 0° - 30° S (SD = 0.33) and the SD of the north equatorial (NE) and south mid-latitude (SM) (SD = 0.18, 0.19 respectively) regions are generally larger than at the poles. This suggests that ridges tend to concentrate at the equator rather than the poles and that ridges preferably formed in the equatorial regions. The SD of the south pole (SP) (SD = 0.12) region is larger than that of the north pole (NP) region (SD = 0.08), which suggests that fluvial erosion and sedimentation and inundation by lakes and seas have reduced the topographic differences in the north polar region in the current season or climatic condition on Titan. In the south polar (SP) region, it is dryer, with fewer lakes and seas, meaning less erosion and sedimentation (or different erosion and sedimentation style), therefore, more ridges preserved and observed by Cassini SAR.
3.4.2 Ridge orientation

The lengths of all mapped ridges per 10° bin of azimuth are totaled and are presented as rose diagrams in Figure 5. Ridge orientations, which are length-weighted by dividing segments of constant orientation into 1 km intervals, are plotted in 30° latitude interval bands and are found in both equatorial and polar regions to be E-W. Notably, in three latitude bands (NE, SE, SM) from 30° N to 60° S, rose diagrams show the scattered trends and highlight a primary dominant orientation (90° azimuth direction) with possible secondary orientation components. Other regions, including polar regions, show only one dominant orientation (90° azimuth direction). This suggests ridges in the equatorial regions have experienced a more complex tectonic history, undergoing possibly more than one tectonic event. The similarity in orientations in the three latitude bands from 30° N to 60° S (NE, SE, SM) suggest those ridges may have formed at the same time and by a common process, where energy was concentrated and stress orientations were systematic (Lopes et al., 2010a; Collins et al., 2009). Systematic contractional or/and extensional stresses through out Titan’s history must have formed the tectonic structures with a preferred orientation (E-W) on Titan’s surface.

In addition, the rose diagram of ridges in the Xanadu region (Figure 5) shows the same tectonic pattern (primary dominant orientation of 90° azimuth with possible secondary components) with the other equatorial regions (NE, SE, SM). This suggests that the ridges in Xanadu may be formed under similar timing with the equatorial tectonic events, which is consistent with the proposed chronology of Titan’s evolution and Xanadu’s geologic history by Langhans et al. (2013). In addition, Langhans et al. (2013) suggest Xanadu was more recently intensely reworked and resurfaced by fluvial processes (also Radebaugh et al. 2011; Burr et al. 2009). Based on our global structural mapping, rose diagrams and structural analysis, the
information may not be sufficient to examine the ancient impact hypothesis for Xanadu formation (Brown et al. 2011). Thus, to understand the origin and geological evolution history of Xanadu province, more detailed regional structural analyses and geodynamic modeling are required.

4. Elevation Distribution

4.1 The elevation of Titan’s ridges

A new topographic map (Lorenz et al., 2013) allows us to examine the correlation between elevation and ridge belts. We overlay our structural mapping (Figure 4) onto the new topographic map (Fig. 3 in Lorenz et al. 2013) (Figure 6). In Figure 6, all mapped structure units are black in order to make the ridges stand out from the topographic map. The tendency of most ridges to lie at higher elevation (red) can be seen. We sorted the ridges into elevation bins of 100 m, ranging from -1500 m to 500 m, where the zero elevation is the 2575 km radius sphere, and compared it with a histogram of Titan’s global topography, following the procedure of Neish and Lorenz (2014). A histogram of Titan’s global topography compared to the elevation of mapped structure units is shown in Figure 7. The distribution of ridges is skewed toward higher elevations. The median elevation of ridges is about 350 m higher than the median elevation of Titan, which suggests that ridge belts preferentially lie at higher elevations.

To investigate this correlation, we used a Kolmogorov–Smirnov test (KS test) which serves as goodness-of-fit technique and tests whether two one-dimensional probability distributions differ (Eadie et al., 1971). The KS test evaluates the statistic D, which quantifies the difference between the cumulative distribution function for Titan’s topography, F(x), and the cumulative distribution function for Titan’s ridge elevations, Fn(x), at elevation x:
\[ D = \sup_x |F_{n}(x) - F(x)| \]

where \( \sup_x \) is the supremum (the least upper bound) of the set of distributions. \( P \) is the distribution function of the KS statistic. Here, the null hypothesis is that the elevations of Titan’s ridges have been drawn at random from the distribution function for Titan’s global topography. We used this technique to determine whether the elevation of Titan’s ridges have been drawn at random from the distribution function for Titan’s global topography (the null hypothesis). For the global topographic data set (Figure 7), this hypothesis has a significance of \( D = 0.325 \), and \( p \) value = 0.02. Thus, the null hypothesis can be rejected with 98% confidence, confirming the conclusion that ridge belts lie at higher elevations.

4.2 Latitudinal distributions of ridge elevations

Here we examine the latitudinal distribution of ridge elevations to see if the ridges tend to lie at higher elevations at all latitudes or only at specific latitudes. We split the map of Titan into six separate latitude bands, similar to the bands chosen for ridge orientations (Section 3), only divided at the equator and each of equal area: (1) 90°N to 42°N, (2) 42°N to 19°N, (3) 19°N to 0°, (4) 0° to 19°S, (5) 19°S to 42°S, (6) 42°S to 90°S. Then, we performed the same analysis we had done for the whole map (section 4.1) to generate the histograms, on each of these six equal area latitude bands, comparing the global elevation in that region to the elevation of the ridges. Here we have six histograms of Titan’s global topography compared to the elevation of mapped structure units in six equal-area latitude bands, shown in Figure 8a - Figure 8f.

At the northern pole (Figure 8a), northern mid-latitude (Figure 9b), southern mid-latitude (Figure 9c), and southern pole (Figure 8d), it is obvious that the histograms of topography and ridge elevation are identical, meaning the ridges do not lie at higher elevations. In contrast, at the
equatorial regions (Figure 8c and Figure 8f), the distribution of ridges is skewed toward higher elevations. Notably, in Figure 8f (0° to 19°S), the peak in lower elevation (-250 m) is likely a result of the lower-elevation rugged Xanadu province. These results show that ridges tend to lie at higher elevation at the equator but not the poles.

5. Discussion and implications for tectonic evolution of Titan

Our structural analysis suggests that the orientation of the ridges on Titan is E-W in both equatorial and polar regions. This result strongly favors a tectonic origin for ridges belts on Titan. Ridge orientations related to impact should show a random or radial rose diagram pattern, which does not agree with the exogenic hypothesis discussed in Moore and Pappalardo (2011).

Mitri et al. (2010) proposes a thermal model resulting in contractional tectonism and suggests the mountains on Titan may be folds. Based on our structural mapping, most of the ridges show linear-to-arcuate morphology similar to the morphology of terrestrial fold-thrust belts, such as Yakima fold-thrust belts. In Figure 9, the morphological similarity of Titan’s equatorial ridge belts and Yakima fold belts is evident. In addition, the average surface slope of Titan’s equatorial ridge belts obtained from SARTopo profiles is < 2° (Liu et al., 2013), which are relatively gentle. Most of the average surface slopes of thrust faults and folds on terrestrial planets are commonly < 15° and the slopes of normal faults are commonly steeper (> 45°). Thus, the combination of these observations and current Cassini data suggests the ridges on Titan are likely folds and thrusts. Higher-resolution images and topographic data would help confirm the origin of the ridges on Titan, for example by locating faults or offset, though much of Titan’s surface is covered in erosional debris.
If we assume the ridges on Titan are folds and thrusts, based on the global contraction models proposed by Beuthe (2010), who used thin elastic shells with variable thickness and suggested the equatorial thinning of the lithosphere would transform the homogeneous and isotropic fault pattern caused by contraction into a pattern of faults striking E-W, preferably formed at the equatorial region (Fig. 8 in Beuthe 2010). If contraction is added to despinning, the despinning pattern first shifts to thrust faults striking N-S and then to thrust faults striking E-W. If the lithosphere is thinner at the poles, the tectonic pattern caused by expansion consists of normal faults striking N-S (Fig. 8 in Beuthe 2010). In addition, the tectonic pattern caused by despinning only would consist of the strike-slip faults in the equatorial region (Fig. 7.4 in Collins et al., 2009). Our mapping shows that the ridge orientation is E-W globally. Analysis of structural density (SD) indicates that ridges are preferably located in the equatorial region. Thus, based on the combination of our mapping, the analysis of rose diagrams and structure density, the most plausible mechanism to form ridges on Titan is ‘contraction dominant’ without the despinning component (Beuthe 2010). If the ridges are not thrusts in nature but normal faults or strike-slip faults, the ridge-forming mechanism may be expansion dominant or despinning.

The analysis of latitudinal distributions of ridge elevations suggests that ridges tend to lie at higher elevation at the equator but not the poles. This may imply that fluvial erosion and sedimentation at the polar regions may fill in the lowlands and flatten the topography, thus, the histograms (Figure 8a and Figure 8d) of topography compared to polar ridge elevations do not show the peak in higher than average elevation.

In sum, two main observations in this study are that ridge belts are E-W orientation globally and that they preferentially lie at higher-than-average elevations in equatorial region. We explore several scenarios that could explain these correlations (Table 2).
Contraction is capable of building ridges at higher elevations and thickening the lithosphere, which is consistent with the observation that most of Earth’s mountains lie in contractional regions at higher elevations, such as the Himalaya and Zagros mountains. In addition, contraction with large localized strain through volume change (Mitri et al., 2010) and/or contraction with lithospheric thinning at the equator (Beuthe 2010) on Titan’s ice shell are capable of generating the elevated ridges striking E-W. A convective upwelling or plume is also capable of building ridges at higher elevations upon the lithosphere. Plumes are postulated to rise through the mantle and begin to partially melt on reaching shallow depths in the asthenosphere by decompression melting. This would create large volumes of magma. On icy satellites, convective plumes cause the melt rise to the surface and erupt to form cryovolcanism (Kargel 1995; Lopes et al., 2010b). However, topographic ridges related to upwelling plumes are typically extensional features, such as rifts, broad topographic rises, and radial fractures which have not commonly been identified on Titan. In addition, the orientation of these extensional features is more likely to have radial patterns. Thus, upwelling plumes or cryovolcanism fail to explain the observations.

Fluvial erosion can create ridges between valleys as seen in the lowlands near Titan’s poles (Barnes et al., 2007). However, pure fluvial erosion would not create ridges with consistent, linear, unidirectional or dual-directional orientations as observed nor explain the location of ridges at higher elevations. In addition, fluvial erosion is not capable of fully eliminating the ridges (Black et al., 2012). Fluvial sedimentation is capable of burying ridges in lowlands but sedimentation would not create ridges with consistent linear E-W orientations. As for aeolian erosion and sedimentation, summative wind directions are though to be westerly near the equator when the greatest transportive/erosive strengths are found (Tokano 2010), consistent
with the orientations of ridges. However, the ridges are globally oriented E-W, even at higher latitudes where winds are not expected to be strong. Additionally, there are equatorial locations where dune orientations differ up to tens of degrees from mountain ridge orientations, so it is not likely that wind erosively formed the ridges. The abundance of ridges at high elevation could be the result of burial of a global population by aeolian sediment at lower-elevation regions, which are mostly at high latitudes. However, most sand seas lie in equatorial highlands instead, where the most dune sedimentation is occurring (Radebaugh et al., 2008; Savage et al., 2013). Thus, pure fluvial and aeolian erosion and sedimentation fail to explain the observations.

Overall, the most reasonable interpretation is that contractional tectonism built the ridges and thickened the icy lithosphere causing regional uplift. Other interpretations, such as upwelling plumes, cryovolcanic rises, fluvial erosion, and aeolian infill, fail to explain all of the observations. In addition, the observation that ridges tend to lie at higher elevation is only true at the equator. Thus, these observations would suggest a surface landscape evolution in which contractional tectonism has built ridges striking dominantly E-W and created high (mainly at the equator) and low (mainly at the poles) topography, and then, more recently, fluvial erosion and sedimentation with seasonal and climatic inputs reduced the topographic differences by filling in the lowlands at high latitudes. The step-by-step scenarios are shown in Figure 10. The interplay of contractional tectonism, erosional and sedimentary processes have created and shaped the current topography we see today.

6. Summary

Global structural analysis leads us to conclude that the origin of most ridge belts on Titan is endogenic, formed by tectonic stresses. This conclusion is consistent with thermal modeling
by Mitri et al. (2010), which showed that several sets of Titan’s ridges could have been formed by contractional tectonism through volume change of Titan’s interior. The orientation of ridges at all latitudes on Titan is E-W. The ridges tend to concentrate in the equatorial regions, which suggests that tectonic activities were more intense near the equator, and that there was possibly more than one tectonic event. The Xanadu province shows an overall similar tectonic pattern with other equatorial regions. The gently arcuate morphology of ridges with gentle surface slopes suggests the ridges are likely contractional structures, i.e. thrusts and folds. If so, the combination of observations suggests that the possible mechanism to produce the E-W ridges on Titan is contraction dominant, with lithospheric thinning at equator (Beuthe 2010).

Using a new topographic map of Titan, we find that Titan’s ridges preferentially lie at higher-than-average elevations. This correlation is only true for the equatorial regions. We explore several explanations for this observed behavior, and judge the most reasonable explanation is that contractional tectonism built the ridges globally, and then fluvial erosion and sedimentation near poles followed, which contributed to shaping Titan’s tectonic landforms and surface morphology into what we see today. Seasonal and climatic changes may affect the distribution of the ridges. This study of structural mapping and distribution analysis can provide constraints on surficial, geological, crustal, and interior evolution and may be tested by more sophisticated contractional or geodynamic models for Titan.
7. References


8. Figures and Captions

Figure 1| Cassini Synthetic Aperture RADAR (SAR) images of Titan’s SAR-bright surface features. (a) The SAR T61 flyby swath (207° W, 10° S) shows SAR-bright ridges as long and curvilinear ridges. SAR-dark, narrow, linear features are eolian dunes. Dunes are distributed on the surface and stop at the elevated ridge regions. (b) Rugged ridges and eroded highlands from the SAR flyby T7 swath. The rugged morphology is different from (a). The curvilinear trend of these rugged ridges is still recognizable even with the eroded morphology. (c) SAR T17 flyby swath (42° W, 8° N) shows three isolated, elevated blocks within the dune-rich area. Dunes diverge around these SAR-bright blocks. (d) SAR T16 swath (142° W, 24° N) demonstrates how the highly curved morphology of mountains on a crater rim differs from that of the more gently curved linear ridges in (a). Image courtesy of NASA/Cassini, north in all images is up.
Figure 2 | (a) The figure demonstrates how ridge crests and basal scarps were defined and mapped. If the ridge appeared to be symmetrical in profile, the ridge crest was mapped. If the ridge was asymmetrical, the basal scarp was mapped with the teeth symbol. The teeth direction is in the possible dip direction of fault plane. (b) The table shows the classification for seven classes of mapped features and their symbols.
Figure 2: (c) The SAR image mosaic shows examples of the long asymmetric ridge scarps (red line with teeth symbol; class 1), the long symmetric ridge crests (red line; class 2), the segmented asymmetric ridge scarps (purple line with teeth symbol; class 3), and the segmented symmetric ridge crests (purple line; class 4). The tear-drop shapes are likely the result of eolian deposits around the ridge. (d) Examples of the eroded, asymmetric ridge scarps (yellow line with teeth; class 5), symmetric ridge crests (yellow line; class 6) and dark, linear valleys (green line) in the Xanadu region. SAR image mosaic in (c) from T8, T61, and T43, SAR image in (d) from T13; courtesy of NASA/Cassini.
**Figure 3** Geometry of SAR image acquisition is illustrated. The depression angle is complementary to look angle; the incidence angle may be affected by planetary curvature. This figure is modified from Ford et al. (2013).
Figure 4] (a) Global structural map of Titan with Cassini SAR images overlain on a background image from the Cassini Visual and Infrared Mapping Spectrometer (VIMS). (b) Titan structural map with Cassini SAR footprint background. Each line is a traced strike of an individual ridge (see the discussion in Section 3.1). Basemap courtesy NASA/ JPL/ USGS.
Figure 5| Lineament azimuth distributions weighted by length. Rose diagrams are plotted in six 30° latitude interval bands: (1) North pole (NP): 90° N - 60° N, (2) North mid-latitude (NM): 60° N - 30° N, (3) North equator (NE): 30° N - 0°, (4) South equator (SE): 0° - 30° S, (5) South mid-latitude (SM): 30° S - 60° S, (6) South pole (SP): 60° S - 90° S. Ridge orientations are length-weighted by dividing segments of constant orientation into 1 km intervals. Rose diagrams of all mapped structures across Titan’s surface, including the Xanadu region (0° - 30° S, 60°W - 150° W), are plotted. One dominant orientation (90° azimuth) can be seen with a possible secondary trend between 30° N to 60° S (NE, SE, SM). These three rose diagram patterns are more scattered compared to other latitude bands. Other regions (NP, NM, SP) show clear one dominant orientation (90° azimuth) only. See Section 3.4.2.

Figure 6| Titan’s ridge belts (black lines) plotted on a global topographic map (Lorenz et al., 2013). The ridges are concentrated in topographically high regions. Cassini SAR swaths are shown shaded.
Figure 7 | Histogram of Titan’s global topography (black dashed line) separated into 100 m bins compared to the elevations of Titan’s ridges (red solid line). Note the tendency for ridges to lie at elevations greater than average.
Figure 8] Histograms of Titan’s global topography (black dashed line) separated into 100 m bins compared to the elevation of Titan’s ridges (red solid line) for each equal latitude band: (a) 90°N-42°N, (b) 42°N-19°N, (c) 19°N-0°, (d) 42°S-90°S, (e) 19°S-42°S and (f) 0°-19°S. The ridges in polar regions do not tend to lie at high elevations, instead, only equatorial regions show the tendency that ridges lie at high elevations. The ridge peak in (c) is obvious. The ridge peak in (f) represents the lower-elevation Xanadu region.
Figure 9] The morphological similarity of Titan’s mountain ridges (left) to the Yakima fold belts in Washington, US (right). The linear-to-arcuate morphology of Titan’s ridge belts and low surface slope suggest they are possibly fold-and-thrust belts comparable to their terrestrial counterparts. SAR image mosaic in (a) from T8, T61, and T43; courtesy of NASA/Cassini. Image in (b) of Yakima fold belts courtesy of Google Earth.
Figure 10 | Model for formation of Titan’s mountain ridges and surface evolution. (a) The schematic diagram shows that contraction thickened the crust and built ridges at higher elevations. (b) Fluvial erosion and deposition followed and filled in the lowlands at higher elevations, which means the ridges lie at higher elevations at the equator only. (c) Finally, the red line represents current, broadly averaged topography (individual ridges are not included). Gray areas represent sediments; black triangles represent ridges.
9. Tables

Table 1| Structure density (SD) data of mapped tectonic features.

<table>
<thead>
<tr>
<th>Region</th>
<th>Latitude (°)</th>
<th>Total SAR area (km²)</th>
<th>SAR coverage (%)</th>
<th>Total structure length (km)</th>
<th>Structure density (SD) (100 km⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>NP</td>
<td>90°N-60°N</td>
<td>3,868,779</td>
<td>69 %</td>
<td>3,085</td>
<td>0.08</td>
</tr>
<tr>
<td>NM</td>
<td>60°N-30°N</td>
<td>7,569,966</td>
<td>50 %</td>
<td>8,662</td>
<td>0.11</td>
</tr>
<tr>
<td>NE</td>
<td>30°N-0°</td>
<td>12,519,518</td>
<td>60 %</td>
<td>20,298</td>
<td>0.18</td>
</tr>
<tr>
<td>SE</td>
<td>0°-30°S</td>
<td>11,453,532</td>
<td>55 %</td>
<td>37,357</td>
<td>0.33</td>
</tr>
<tr>
<td>SM</td>
<td>30°S-60°S</td>
<td>6,114,171</td>
<td>40 %</td>
<td>11,399</td>
<td>0.19</td>
</tr>
<tr>
<td>SP</td>
<td>60°S-90°S</td>
<td>3,823,641</td>
<td>68 %</td>
<td>4,587</td>
<td>0.12</td>
</tr>
</tbody>
</table>

Table 2| Summary of the scenarios that could explain the correlation between ridge distribution and elevation on Titan.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Pros</th>
<th>Cons</th>
</tr>
</thead>
<tbody>
<tr>
<td>Global contraction</td>
<td>• Capable of building ridges at higher elevation</td>
<td>• Global contraction would not build the ridges with preferred orientation</td>
</tr>
<tr>
<td></td>
<td>• Consistent with the observation that most of Earth’s mountains lie at higher elevations</td>
<td></td>
</tr>
<tr>
<td>Upwelling plume</td>
<td>• Capable of building ridges at higher elevations</td>
<td>• Ridges built by upwelling plume are extensional features, such as rifts, broad topographic rise, shield volcanoes and radial features, which are not seen on the surface of Titan</td>
</tr>
<tr>
<td></td>
<td></td>
<td>• Cryvolcanism is not prevalent</td>
</tr>
<tr>
<td></td>
<td></td>
<td>• Cannot explain the ridges in equatorial highlands only by pure fluvial erosion since fluvial erosion would not create preferred-orientation ridges</td>
</tr>
<tr>
<td>Fluvial erosion/</td>
<td>• Capable of eroding ridges at the lowlands at high latitudes</td>
<td></td>
</tr>
<tr>
<td>sedimentation</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Aeolian infill</td>
<td>• Capable of burying ridges at the lowlands</td>
<td>• Most sand seas lie at equatorial highlands</td>
</tr>
</tbody>
</table>
Chapter 2:

Role of Fluids in Mechanics of Overthrusting Faulting on Titan


The general understanding of icy satellite tectonics prior to the arrival of the Cassini spacecraft at Saturn and its largest moon Titan (2004- present) was that most exhibit evidence for extensional tectonism (e.g. fractures, grabens, normal faults), whereas evidence for contractional tectonism (e.g. folds, thrust faults) is rare (Collins et al., 2009). The Cassini RADAR instrument, operating in Synthetic Aperture RADAR (SAR) mode, has obtained images at ~350 m resolution of many landforms on Titan (Elachi et al., 2005; Lunine et al., 2008), including E-W oriented, long, narrow mountain ridges, near the equator, termed ridge belts (Radebaugh et al., 2007; Liu et al., 2014) (Figure 1). Several studies of the ridge belts have concluded that they formed by contractional tectonism. Their long, curvilinear morphology (Liu et al., 2012; Radebaugh et al., 2011; Solomonidou et al., 2013), their low slopes and topography (Radebaugh et al., 2007; Liu et al., 2012; Mitri et al., 2010a), comparisons of these morphologies with Earth’s tectonic features (Solomonidou et al., 2013), and structural and stress field analysis (Liu et al., 2012; Liu et al., 2014) are all consistent with formation by contraction (Figure 1). If this is true, Titan may be the only icy satellite on which widespread contraction has occurred, and, in fact, on which contraction may be the predominant style of tectonism (Mitri et al., 2010). However, the differential stress required for contractional strain is considerably greater than for extensional strain. For example, the stress required to form contractional structures on the icy satellites of Jupiter, Ganymede and Europa, is 10-25 MPa. This is 3-8 times that required to form extensional...
features (Pappalardo and Davis 2007; Dombard and McKinnon 2006). Generating such large stresses on icy satellites is difficult. The dominant source of stress for brittle-frictional faulting, which is diurnal eccentricity, can only generate stresses of 0.1 MPa on Europa (Collins et al., 2009) and smaller than the order of 0.1 MPa on Ganymede and Titan (Collins et al., 2009). Other mechanisms may produce larger stresses, such as despinning, non-synchronous rotation and polar wander (Dombard and McKinnon 2006). However, it is unclear if these mechanisms have been involved in Titan’s tectonic history. In summary, sources of contractional stress sufficient for thrust faulting probably do not exist on most icy satellites, including Titan. Thus, a paradox has emerged, wherein no stress source is known that is large enough to produce the contractional structures observed on Titan.

In this study, we provide a solution for the stress paradox on Titan: fluid pressure. Liquid hydrocarbons (e.g. methane, ethane) have been identified on Titan’s surface (Stofan et al., 2007; Lorenz et al., 2008) and may flow in the subsurface (Hayes et al., 2008; Tomasko et al., 2005). Therefore, Titan has a hydrologic cycle that likely includes a ground “methane” system similar to Earth’s groundwater system (Lunine and Lorenz 2009). Here we report that fluid pressures associated with liquid hydrocarbons in Titan’s subsurface significantly reduce the shear strength of the icy crust and enable contractional structures to form without the requirement of large stresses. In this paper, we first estimate the strength of Titan’s icy lithosphere (the absolute differential stress necessary for failure) in the brittle and ductile regimes, and then we discuss the role of fluid pore pressure on Earth and how it reduces the stress needed for contraction, and finally discuss its application to Titan’s unique crustal environment.

The strength of the lithosphere depends on its composition, pressure, temperature, and strain rate. In the upper lithosphere, strength is primarily controlled by increasing lithostatic
pressure with depth, according to Byerlee’s law (Byerlee 1978). As temperature increases in the lower part of the lithosphere, strength decreases with depth and is controlled by ductile flow. The composition of Titan’s crust is mainly water ice, perhaps mixed with methane hydrate at shallow depth (Lunine and Lorenz 2009). Here, we adapt the behavior of ice based on Byerlee’s law to estimate the conditions for brittle failure of Titan’s upper lithosphere, assuming a friction coefficient \( \mu = 0.69 \) (Byerlee 1978; Beeman et al., 1988). The yield stress envelope for this case (Watts 2001; Luttrell and Sandwell 2006) is:

\[
\sigma_c = k \rho g z
\]

where \( \sigma_c \) is the stress for contraction, \( \rho \) is the density of ice, \( g \) is the acceleration of gravity, \( z \) is the depth, and the dimensionless factor \( k = 2.6 \) is a constant from the application of Byerlee’s law to ice (Watts 2001). The crustal parameters of Titan are summarized in Table 1.

The behavior of the lower part of Titan’s lithosphere, that controlled by ductile flow, is assumed to follow the constitutive rheology of viscous flow as based on laboratory measurements (Durham et al., 1997; Goldsby and Kohlstedt 2001). The rheological parameters of ice used in this work are summarized in Table 2. The strain rate \( \varepsilon \) is based on the flow law (Mitri et al., 2010a):

\[
\varepsilon = A \left( \frac{1}{d} \right)^m \sigma^n \exp \left( -\frac{Q}{RT} \right)
\]

where \( A \) is a material-dependent constant, \( d \) is the grain size, \( m \) is the grain-size exponent, \( n \) is power-law exponent, \( Q \) is the activation energy, \( R \) is the gas constant, and \( T \) is the absolute surface temperature. The rheology of ductile creep in regime B is adopted (Pappalardo and Davis 2007) here, based on the modeled ice shell temperature range (< 240 K) of Titan (Mitri et al.,
We assume a surface temperature of 94 K (Tobie et al., 2005), and a thermal gradient of 5 K km\(^{-1}\) (Jaumann et al., 2010). The ice grain size is not well constrained, so we adopt a grain size of 1 mm to be consistent with the current literature for icy satellite tectonics (Barr and McKinnon 2007).

Figure 2 illustrates Titan’s lithospheric strength as a function of depth in the brittle and ductile regimes. The black straight line represents the failure strength of the pre-fractured and frictionally controlled brittle lithosphere in contraction from equation (1). The blue curves represent strain rates of 10\(^{-14}\), 10\(^{-15}\), and 10\(^{-16}\) s\(^{-1}\) for ice in a ductile B creep regime and are calculated by equation (2). The maximum lithospheric strength is approximated by the intersection of the brittle and ductile failure curves, suggesting Titan’s lithospheric strength in contraction is \(~8\text{-}10\) MPa, which is comparable to estimates for Ganymede’s lithospheric strength (\(~11\) MPa) (Pappalardo and Davis 2007). In other words, to enable contractional strain on Titan, a stress of less than 8-10 MPa is required.

On Earth, some thrust faults exist that required horizontal motions of thrust sheets over distances of as much as 100 km. However, the stresses needed to move the thrust sheets exceed the crushing strength of rocks, therefore yielding a strength paradox. This was resolved by Hubbert and Rubey (1959) who demonstrated that high fluid pressures reduce the normal stress along a fault plane, thereby significantly reducing frictional resistance to sliding. Understanding this solution to the paradox involved modifying the Mohr-Coulomb law of frictional shear strength:

\[
\tau = \mu (\sigma_N - P_f) + C
\]  

(3)
where $\tau$ is the shear stress required to cause slip, $C$ is cohesion, $\mu$ is the coefficient of internal friction, $\sigma_N$ is normal stress, and $P_f$ is fluid pore pressure. Where fluid is unable to escape from a porous but impermeable subsurface horizon, fluid pressure ($P_f$) can build to the point of supporting most of the weight of the overlying rock. These conditions significantly reduce the frictional resistance to sliding along the overpressured horizon to near zero. In other words, the fluid pressure is high enough to offset nearly all the normal stress ($\sigma_N$). Consequently, the shear stress ($\tau$) needed to form a thrust fault must only exceed the cohesive strength ($C$) of rock. Under these conditions a small convergence can easily overcome the sliding friction to form thrust sheets and related folds.

Consider a porous rock with a strength envelope represented by a Mohr diagram (Figure 3). This rock is subject to an initial stress condition that is stable (blue Mohr circle in Figure 3). The subsequent addition of fluid pressure has the effect of lowering the principal stresses, equally in all directions because it is hydrostatic. The Mohr circle remains the same size and it moves to the left on the horizontal axis, where it intersects the failure envelope. Thus, the addition of pore fluid causes the rock to fracture.

Titan’s upper icy crust is likely porous, fractured from impacts, may contain methane clathrate hydrates (Lunine and Lorenz 2009; Janssen et al., 2009), and is possibly filled with hydrocarbon fluid in a hydrocarbon-based hydrologic system (Lunine and Lorenz 2009). We adapt the Mohr-Coulomb equation (3) for Titan and assume pore fluid pressures ($P_f$) on Titan are locally high enough to offset all normal stresses ($\sigma_N$) as they appear to be on Earth. The cohesive strength ($C$) of Titan’s fractured, icy crust may be about 1 MPa or smaller (Beeman et al., 1988). Based on equation (3), $\sigma_N - P_f = 0$, resulting in $\tau = C$. Thus, to enable brittle contractional strain on Titan, large stresses (8~10 MPa) are not necessary if the crust contains trapped pore fluids.
Only stresses ($\tau$) of 1 MPa or smaller may be required to overcome the cohesion of ice and form fold and thrust belts on Titan. We use the Mohr circle (Figure 3) to demonstrate the effects of fluids on the crustal ice-methane hydrate mixture. The pore pressure caused by hydrocarbon fluids on Titan would have the same effect on the ice mixture to move the Mohr circle to the left along the horizontal axis and to intersect the failure envelope.

Another important aspect of crustal fluids in Titan is the pore fluid pressure ratio ($\lambda$). Hydrostatic pressure in an aquifer corresponding to a pore fluid pressure ratio ($\lambda$) is defined as $\lambda = \rho_f/\rho_s$, also called the fluid pressure gradient (Davis et al., 1983), where $\rho_f$ is the density of the fluid and $\rho_s$ is the density of the crust. We can revise equation (3) in terms of the fluid pressure gradient ($\lambda$) (modified from Davis et al. 1983):

$$\tau = C + \mu\sigma_N^* = C + \mu\sigma_N (1 - \lambda)$$  \hspace{1cm} (4)

where $\sigma_N^*$ is the effective normal stress fraction, the fractional value for normal stress remaining after it has been reduced by fluid pore pressure. On Earth, with a fluid (liquid water) density of 1 g/cm$^3$ and a crustal density of 2.5 g/cm$^3$, the hydrostatic pore fluid pressure ratio $\lambda$ is ~0.4. Based on equation (4), hydrostatic pore pressure reduces the normal stress, leaving an effective normal stress fraction ($\sigma_N^*$) ~60% of the original value. On Titan, with a fluid (liquid methane) density of 0.6 g/cm$^3$ (Lorenz 2002) and an icy crust density of 0.9 g/cm$^3$ (Tobie et al., 2005), the pore fluid pressure ratio $\lambda$ is ~0.67, about 1.5 times greater than that for rocks on Earth. Thus, hydrostatic pore pressure on Titan reduces the normal stress even more than on Earth, leaving an effective normal stress fraction ($\sigma_N^*$) ~33% of the original value. On Titan, where crustal conditions create a higher hydrostatic fluid pressure gradient ($\lambda$), there is a greater reduction in
shear stress needed to enact brittle strain, which makes formation of contractional structures even easier on Titan than on Earth.

Below most terrestrial fold-and-thrust belts, if permeability is restricted and fluid is trapped, pore pressures may exceed the hydrostatic pressure and the fluid becomes overpressured. The fluid pressure gradient may exceed the hydrostatic fluid pressure ratio ($\lambda > 0.4$ for Earth and $\lambda > 0.67$ for Titan). In Earth’s crust, it is common to have a fluid pressure ratio $\lambda > 0.4$ (Davis et al., 1983). Thus, pore fluid pressure ratios below Titan’s fold belts may also exceed the hydrostatic pressure and reach even higher values of $\lambda$. In addition, Mousis and Schmitt (2008) suggest the presence of clathrates of methane and water within the crust reduces its permeability by closing pore space networks connecting to the surface. This decrease in permeability would increase pore fluid pressures, reducing the shear strength of Titan’s crust. However, experimental studies are needed to help us understand how Titan’s icy crust and hydrocarbon liquids chemically interact and the effect of these mixtures on crustal strength.

In Figure 2, the red dashed line represents the failure strength of Titan’s lithosphere with a hydrostatic fluid pore pressure ($\lambda \sim 0.67$). The hydrostatic fluid pore pressure reduces the maximum absolute differential stress needed to enact contractional failure to ~3-5 MPa. The green dashed line demonstrates that with higher pore fluid pressure ($\lambda \sim 0.9$), the stress needed to form contraction decreases significantly, to ~1-2 MPa. Thus, the fluid pressures associated with liquid hydrocarbons in Titan’s subsurface may significantly reduce the shear strength of the icy crust and enable contractional faults to form without requiring large stresses.

Since the hydrocarbon liquids are likely located in the shallow part of the icy crust (2-3 km) (Lunine and Lorenz 2009; Choukroun et al., 2010), faults and associated crustal deformation
do not need to involve the entire thickness of the icy crust. Thus, the proper conditions may exist for thin-skinned tectonics. In this style of tectonic deformation, when a crustal section with fluid overpressures is subjected to a large enough horizontal compressive stress, it deforms internally and forms low-angle thrust faults and folds along a basal surface (décollement).

The lower boundary of the décollement depth is approximately \( w \tan \theta \), where \( w \) is the fold width, \( \theta \) is the wedge angle \( = \alpha + \beta \), \( \alpha \) is the surface slope and \( \beta \) is the décollement dip angle (Davis et al., 1983). We obtain the fold width from Cassini SAR images by measuring the width of a mountain ridge belt, which is visible as a SAR-bright linear trough because they are rough, fractured, and water-ice-rich, in contrast with surrounding, SAR-dark, smooth or fine-grained and sandy terrains (Radebaugh et al., 2007; Liu et al., 2012). Beginning at a bright/dark boundary, we measured the average surface slope of fold belts using the SARTopo technique (Stiles et al., 2009), which obtains topography from overlapping SAR images (See Supplementary Discussion S1). For the fold-and-thrust belts located at 200°W, 10°S on Titan (Figure 4), the surface slope \( \alpha \) is \( \sim 1.5°-2° \) (See Supplementary Figure S1 and Table S1). The width of a single fold is \( \sim 30-45 \) km. The décollement dip angle (\( \beta \)) cannot be obtained with the current Cassini topography data. On Earth, \( 0° < \beta < 10° \) is considered to be a reasonable range (Davis et al., 1983). Here we adopt this range of dip angles to estimate the décollement depth. Using the estimated values of \( w \), \( \alpha \) and \( \beta \), the lower boundary of the décollement depth is approximately \( \sim 2 \) to \( 3 \) km. Using the topographic profiles of the fold belt, the estimate of décollement depth, and the assumption of high pore fluid pressure with thin-skinned tectonics, we are able to construct a generalized cross-section of Titan’s fold belts at 200°W, 10°S (Figure 4). In the thin-skinned model, the upper crust is detached from the underlying basement along
fault planes with a ramp-flat geometry; thus, the thrusting angle should be higher than the overall slope angle of the fold belts.

Titan is the only icy satellite on which there is strong morphological evidence for contractional deformation. What makes Titan unique? Mitri et al. (2010) suggest that icy satellites without high-pressure phases of water ice layers (e.g. the ice I shell of Europa) experience global extension due to thickening of the icy crust during progressive cooling. In contrast, Titan is larger and it should have developed high-pressure ices, which would lead to global contraction during internal cooling. However, Ganymede and Callisto, which also should have high-pressure ice, do not show strong evidence for contractional deformation (Mitri et al., 2010). We therefore conclude that high-pressure phases of ice are not sufficient to cause contractional features. An overpressured crustal fluid may also be required. Among the icy satellites, this pair of conditions —high pressure ice phases inducing contraction and a low density crustal fluid—is only met on Titan.

Titan’s formation at greater distances from the Sun and at lower temperatures allowed for the incorporation of significant amounts of ammonia and methane ices – materials more volatile than water ice when compared with Jupiter’s moons. Because of this extra volatile endowment, Titan has a thick, nitrogen-rich atmosphere, which, because of its density (and the ambient temperature) stabilized liquid hydrocarbons and established a hydrocarbon-based hydrologic system complete with liquids flowing on and below its surface. Thus, through geological time, Titan’s thick atmosphere has sustained the stability of surface and subsurface liquids, weakening the upper crust. On the other hand, Callisto and Ganymede lack significant proportions of these volatiles, and hence lack atmospheres and hydrospheres with ground-liquids. Consequently, they have very different tectonic systems than Titan and they lack contractional tectonic features.
Thus, the volatile component obtained during accretion directly affects the style of tectonics on icy satellites, just as it does on Earth.

The facilitation of thrust faults on Titan through the presence of hydrocarbon liquids and resultant crustal weakening yields contractional features—fold and thrust belts—with morphologies similar to those on Earth (Radebaugh et al., 2007; Mitri et al., 2010). This similarity strengthens the case for ongoing comparisons of surface and near-surface processes on Titan and Earth. This study furthermore advances the understanding of how liquids, other than water, play an essential role in the tectonic evolution of icy satellites. Finally, our conclusions highlight the significance of fluids in planetary lithospheres and the importance of atmospheres and initial compositions for the development of planetary tectonic systems. This has implications for tectonic processes on all solid planetary bodies that may have fluid in their lithospheres, now or in the past.
References


**Figure 1** | **Mountain ridge belts on Titan.** A Cassini SAR (Synthetic Aperture Radar) mosaic (T8, T61 and T41 flyby swaths) with an Imaging Science Subsystem (ISS) image as the background shows a series of bright linear ridges at 200°W, 10°S on Titan. SAR image brightness represents the normalized microwave energy backscattered from the surface, which is a function of surface slope, dielectric properties, roughness, and the amount of volume scattering. SAR images cover ~55% of Titan surface to date and have a resolution of ~350 m/pixel. Ridge belts are elevated, SAR-bright features with curvilinear margins morphologically similar to terrestrial fold belts. Red rectangle shows the location of the image in Figure 4.
Figure 2 | Titan’s lithospheric strength as a function of depth in the brittle and ductile regimes. Brittle failure lines (black, red, and green) show lithospheric strength for contraction under different conditions. The black line represents the strength with no fluid effects ($\lambda = 0$); the red dashed line is for moderate hydrostatic fluid pressure ratio ($\lambda = 0.67$); and the green dashed line is for a high fluid pressure ratio ($\lambda = 0.9$). Ductile strength curves (in blue) for ice use the ductile creep B regime (Pappalardo and Davis 2007), a surface temperature of 94 K (Tobie et al., 2005), and a thermal gradient of 5 K km$^{-1}$ (Jaumann et al., 2010), at various strain rates ($10^{-14}$, $10^{-15}$, and $10^{-16}$ s$^{-1}$). The maximum lithospheric strength is approximated by the intersection of a brittle and ductile failure curve. Thus, the maximum lithospheric strength for a “dry” lithosphere and a rapid strain rate of $10^{-14}$ s$^{-1}$ is about 10 MPa; for a high pore fluid pressure and low strain rate, the strength is about 1.3 MPa.
Figure 3 | Mohr circle plot and the fluid effect. Mohr diagram shows the effect of fluid pore pressure on brittle failure. $\sigma_n$ is normal stress and $\sigma_s$ is shear stress. Each Mohr circle represents the same state of effective stress. The blue Mohr circle represents the initial stresses with no pore pressure. Pore fluid pressure reduces the normal stress and translates the Mohr circle to the left, as indicated by the red-dashed Mohr circle of effective stress. Once the Mohr circle intersects the failure envelope, the rock fractures.
Figure 4 | The topography and interpreted cross-section of an equatorial ridge belt on Titan. (A) A SARTopo profile (colored line, coded by elevation) (Stiles et al., 2009) is shown on the Cassini T61 flyby swath. Radar bright curvilinear features are ridge belt. Radar black lines are sand dunes, which stop at the higher elevation ridges. (B) In profile a-a’, the black line is the actual elevation extracted from SARTopo data and the red dashed line is the interpreted topography. Vertical exaggeration of topographic profile portion (0-0.4 km) is: 50:1. The short blue lines represent crustal hydrocarbon (methane and ethane) fluids trapped in pore spaces. The gray patches at the surface represent eroded debris and sediment accumulated between the ridges. The interpretive cross section a-a’ assumes that thin-skinned tectonics are appropriate for Titan and shows a series of imbricate thrust faults splay upward from a basal décollement beneath surface folds. Vertical exaggeration of subsurface cross section (0-7 km) is: 5:1.
## Tables

**Table 1** Crustal parameters of Titan.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
<th>Value</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \rho )</td>
<td>Density of lithosphere (Ice I)</td>
<td>900 kg m(^{-3})</td>
<td>Tobie et al. (2005)</td>
</tr>
<tr>
<td>( g )</td>
<td>Acceleration due to gravity</td>
<td>1.352 m s(^{-2})</td>
<td>Iess et al. (2010)</td>
</tr>
<tr>
<td>( \rho_f )</td>
<td>Density of the hydrocarbon fluid</td>
<td>600 kg m(^{-3})</td>
<td>Lorenz (2002)</td>
</tr>
<tr>
<td>( C )</td>
<td>Cohesion</td>
<td>1 MPa</td>
<td>Beeman et al. (1988)</td>
</tr>
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</table>

**Table 2** Rheological parameters and constants used in equation (2).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
<th>Value</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Log ( A )</td>
<td>Material-dependent constant</td>
<td>11.8 MPa(^{m}m^{n}s^{-1})</td>
<td>Durham et al. (1997)</td>
</tr>
<tr>
<td>( m )</td>
<td>Grain-size exponent</td>
<td>0</td>
<td>Durham et al. (1997)</td>
</tr>
<tr>
<td>( n )</td>
<td>Power-law exponent</td>
<td>4.0</td>
<td>Durham et al. (1997)</td>
</tr>
<tr>
<td>( d )</td>
<td>Grain size</td>
<td>1 mm</td>
<td>Barr and McKinnon (2007)</td>
</tr>
<tr>
<td>( Q )</td>
<td>Activation energy</td>
<td>61 KJ mol(^{-1})</td>
<td>Durham et al. (1997)</td>
</tr>
<tr>
<td>( T )</td>
<td>Absolute temperature</td>
<td>94 K</td>
<td>Tobie et al. (2005)</td>
</tr>
<tr>
<td>( R )</td>
<td>Gas constant</td>
<td>8.3144621 J mol(^{-1})K(^{-1})</td>
<td>Durham et al. (1997)</td>
</tr>
</tbody>
</table>
Supplementary Material- Chapter 2

“Role of Fluids in the Mechanics of Thrust Faulting on Titan”

Contents:
1. Supplementary Discussion S1
2. Supplementary Figure S1
3. Supplementary Table S1

1. Supplementary Discussion S1:

Slope measurements on Titan’s fold belts:

Topographic profiles of ridge belts on Titan (Figure S1) were constructed using
SARTopo techniques (Stiles et al., 2009), which obtains absolute topography respect to the 2575
km radius sphere from overlapping SAR images. It enables two or more surface-height profiles
to be obtained from the long dimension of each SAR image. This technique depends upon
accurate knowledge of spacecraft attitude and the antenna gain pattern. The vertical resolution of
SARTopo profiles is 75 meters and the horizontal resolution is 10 km. The SARTopo profiles are
narrow bands, not everywhere normal to ridge belt strikes. Thus, the average slopes (α’) we
obtained from SARTopo profiles must be converted to true slope (α) through the simple
relationship tan α’ = tan α sin δ, where δ is the angle between the line of ridge strike and the line
of SARTopo trace. See Figure S1 and Table S1 for more details on the measurement results.

Since SARTopo data has a vertical uncertainty of 75 m and a horizontal uncertainty of 10
km, an estimate of the surface slope ~ tan⁻¹(ridge height/ ridge width) has an uncertainty of
approximately 0.4°.
2. Supplementary Figure:

![Figure S1](image)

Figure S1 | A SARTopo trace (colored line) is shown on the Cassini SAR T61 flyby swath with a topographic relief profile. Vertical exaggeration: 250: 1. The coordinates of the center of this image are 208.3°W, 3.4°S. α’ is the apparent slope or the average slope obtained from the SARTopo profile and δ is the angle between the line of mountain strike and the line of SARTopo trace. The true slopes α for folds from left to right are 1.8°, 1.8°, 1.9° and 1.5°, respectively. The details of the calculated slope angles for each measurement can be found in Supplementary Table S1.
3. Supplementary Tables:

Table S1 | Slope measurements of fold belts on Titan.

<table>
<thead>
<tr>
<th>Location</th>
<th>SAR flyby</th>
<th>H (m)</th>
<th>Length (km)</th>
<th>α’</th>
<th>δ</th>
<th>α</th>
</tr>
</thead>
<tbody>
<tr>
<td>(207.4°W, 4.0°S)</td>
<td>T61</td>
<td>220</td>
<td>18</td>
<td>0.7°</td>
<td>23°</td>
<td>1.8°</td>
</tr>
<tr>
<td>(207.4°W, 4.0°S)</td>
<td>T61</td>
<td>180</td>
<td>15</td>
<td>0.7°</td>
<td>23°</td>
<td>1.8°</td>
</tr>
<tr>
<td>(210.4°W, 4.6°S)</td>
<td>T61</td>
<td>250</td>
<td>22</td>
<td>0.6°</td>
<td>20°</td>
<td>1.9°</td>
</tr>
<tr>
<td>(210.4°W, 4.6°S)</td>
<td>T61</td>
<td>150</td>
<td>20</td>
<td>0.4°</td>
<td>20°</td>
<td>1.5°</td>
</tr>
</tbody>
</table>

(α error ~±0.4°)
ABSTRACT

Chapter 3: Tsunami Modeling of 1629 Mega-thrust Earthquake in Eastern Indonesia

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Arthur Wichmann’s ‘Earthquakes of the Indian Archipelago’ documents several large earthquakes and tsunami throughout the Banda Arc region that can be interpreted as mega-thrust events. However, the source regions of these events are not known. One of the largest and well-documented events in the catalog is the great earthquake and tsunami affecting the Banda islands on 1 August 1629. It caused severe damage from a 15-meter tsunami that arrived at the Banda Islands about a half hour after violent shaking stopped. The earthquake was also recorded 230 km away in Ambon, but no tsunami is mentioned. This event was followed by at least 9 years of uncommonly frequent seismic activity in the region that tapered off with time, which can be interpreted as aftershocks. The combination of these observations indicates that the earthquake was most likely a mega-thrust event.

We use an inverse modeling approach to numerically reconstruct the tsunami, which constrains the likely location and magnitude of the 1629 earthquake. Only linear numerical models are applied due to the low-resolution of bathymetry in the Banda Islands and Ambon. Therefore, we apply various wave amplification factors (1.5 to 4) derived from simulations of recent, well-constrained tsunami to bracket the upper and lower limits of earthquake moment magnitudes for the event.

The closest major earthquake sources to the Banda Islands are the Tanimbar and Seram Troughs of the Banda subduction/collision zone. Other source regions are too far away for such a short arrival time of the tsunami after shaking. Moment magnitudes predicted by the models in order to produce a 15 m tsunami are Mw of 9.8 to 9.2 on the Tanimbar Trough and Mw 8.8 to 8.2 on the Seram Trough. The arrival times of these waves are 58 minutes for Tanimbar Trough and 30 minutes for Seram Trough. The model also predicts 5 meters run-up for Ambon from a Tanimbar Trough source, which is inconsistent with the historical records. Ambon is mostly shielded from a wave generated by a Seram Trough Source.

We conclude that the most likely source of the 1629 mega-thrust earthquake is the Seram Trough. Only one earthquake > Mw 8.0 is recorded instrumentally from the eastern Indonesia region although high rates of strain (50-80 mm/a) are measured across the Seram section of the Banda subduction zone. Enough strain has already accumulated since the last major historical event to produce an earthquake of similar size to the 1629 event. Due to the rapid population growth in coastal areas in this region, it is imperative that the most vulnerable coastal areas prepare accordingly.

Keywords: Tsunami modeling, Indonesia, Banda arc, Seram Trough, Tanimbar Trough, Banda Islands, Ambon, Mega-thrust earthquakes
Chapter 3:

Tsunami Modeling of 1629 Mega-thrust Earthquake in Eastern Indonesia

1. Introduction

During the twentieth century Indonesia had around two hundred major earthquakes (M\textsubscript{s} 7.5 or greater), more than all of North America or South America during the same time interval (Harris et al. 1997). At least 110 of these quakes were destructive; the majority jolting densely populated western Indonesia (Figure 1).

These high rates of seismic activity in the past century are consistent with the recorded history of Indonesia, as documented in a compilation of geophysical events from the 17\textsuperscript{th} to 19\textsuperscript{th} centuries by Wichmann (1918). The reliability of the earthquake and tsunami catalog is demonstrated by the recent recurrence of several earthquakes that ruptured similar fault segments and are of similar magnitudes to those inferred from earlier accounts. For example, the 2005 northern Sumatra earthquake near Nias Island ruptured nearly the same area estimated by Newcomb and McCann (1987) from accounts in the Wichmann catalog of an earthquake and tsunami in 1861.

The M\textsubscript{w} 9.2 Banda Ache mega-thrust earthquake, and the spatial and temporal clustering of events that followed, have stimulated new interest in the earthquake history of the western Sunda Arc. All eyes are presently on the segment of the Sumatran subduction zone to the south of Nias, which last ruptured in 1833. From reports of this event and the tsunami that followed
that are recorded in the Wichmann catalog Newcomb and McCann, (1987) estimate it could have been up to Mw 9.0.

The Sunda arc-trench system continues to the east of Sumatra adjacent to the densely populated islands of Java and Bali, then becomes the Banda Arc in the Timor region (Figure 1). Although no earthquakes larger than Mw 8.0 are documented in the Java-Bali region, there are several earthquakes in the Banda Arc region documented before 1900 by Wichmann with similar characteristics to those in Sumatra. However, the source parameters of the large events are unknown. No large earthquakes in the Banda Arc and Eastern Indonesia region are included in the recent summary of mega-thrust earthquake events worldwide by Heuret et al (2012), which is based only on the last 100 years of instrumental records. When these largely unknown events reoccur in eastern Indonesia it will affect an order of magnitude more people and urban centers than before (Figure 1), which raises the stakes on determining the most likely large earthquake and tsunami sources in the region. This paper investigates the most likely source of one of the largest tsunami events documented in Eastern Indonesian region, which occurred in 1629.

2. Tectonic Setting

The Banda Arc region occupies a convergent triple junction of three of Earth’s largest plates. Relative to the Sundaland block, a tectonic domain that is extruding eastward away from Eurasian plate (Rangin et al, 1999), the Indo-Australian Plate converges NNE at a rate of around 70 mm/a (Nugroho et al., 2009). The Indo-Australian Plate subducts beneath the Banda Arc from three sides along the Timor, Tanimbar and Seram Troughs. These troughs are underthrust mostly by Australian continental margin lithosphere that is attached to old Indian Ocean lithosphere at deeper levels of the subduction zone (Hamilton, 1979). The resulting arc-continent collision bends 180° around the western edge of NW Australia (Figure 2). Another major
tectonic player in the region is the Pacific Plate. It moves WNW at a rate of around 110 mm/a relative to the Sundaland Plate (Rangin et al, 1999), and side-swipes the northern part of the Banda Arc along an array of plate boundary segments linked by the left-lateral Sorong Fault (Figures 1 and 2).

The NW part of the Australian continental margin first began to subduct beneath the Banda Arc at around 8 Ma in the Timor and Seram regions (Berry and McDougall 1986; Linthout et al., 1996; Harris, 2011). The arc-continent collision has now propagated around the embayment south of the Bird’s Head and involves the entire Banda Arc region (Hall, 2002). Underthrusting of the Australian continental margin increases coupling along the subduction interface as indicated by partitioning of strain away from the deformation front into the forearc (Reed et al., 1986; Harris et al., 2009) and back arc regions (Silver et al., 1983). In the backarc the volcanic arc is thrust over the southern edge of the Banda Sea Basin (Figure 2) along the Flores and Wetar Thrusts (Breen et al., 1989). These south-dipping backarc thrust systems (Figure 1 and 2) currently take-up a significant amount of the convergence between the Australia and Asia plates (McCaffrey and Nableck, 1984; Genrich et al., 1996). The amount of movement along these thrust systems decreases to the west where the oblique collision is less developed (Silver et al., 1983). Even in the most advanced part of the collision in the Timor region there is still up to 21 mm/a of convergence measured across the Timor Trough (Nugroho et al., 2009), which means that a significant amount of elastic strain energy is accumulating along this plate boundary (Harris, 2011).

The low rate of seismicity along the Timor, Tanimbar and Seram Troughs is commonly cited as evidence that they are no longer active (McCaffrey et al., 1985). However, seismic reflection profiles across the deformation front at the Seram, Tanimbar and Timor Troughs show
thrust faults breaking all of the way to the surface (Schluter and Fritsch, 1985; Karig et al., 1987; Pairault et al., 2003). We interpret the low seismic slip rates of the Banda collision zone as evidence of a locked plate boundary interface, which potentially could produce mega-thrust earthquakes. There are several small thrust-mechanism earthquakes along the low-angle plate interface of both the Timor and Seram Troughs. Although active thrusting is observed in earthquakes along the plate boundary in the Tanimbar Trough region there are many extensional and strike-slip events as well (McCaffrey et al., 1985).

The Backarc thrust systems may also produce large earthquakes. The mapped length of the Wetar Thrust is around 350 km, which may be too short for a mega-thrust earthquake event. However, the Flores Thrust is at least 500 km in length, which, if the whole thrust system ruptured, may produce a mega-thrust earthquake.

The working definition here for a mega-thrust event is an inter-plate earthquake that occurs along a subduction interface that is Mw ≥ 8.5 and is associated with uncommonly large rupture lengths (> 400 km). The fault plane commonly has a shallow dip (<10°) and the earthquake a shallow hypocenter (<30 km), which at a subduction zone can produce large tsunami.

Most mega-thrust earthquakes of the magnitude needed to produce the tsunami observed in the Banda Islands have rupture lengths of > 500 km, such as 2010 Mw 8.8 Maule “Chile” earthquake (Wang et al. 2012) and 2011 Mw 9.0 Tohoku earthquake (Suzuki et al.2011). However, the most significant problem with either of the backarc thrust systems sourcing the 1629 tsunami is their distance (> 450 km) from the Banda Islands where the tsunami arrived only 30 minutes after intense shaking (See Section 2.1 below).
It is also possible that unknown active faults exist close enough to the Banda Islands to generate a tsunami that arrives within 30 minutes. For example, active strike-slip faults are inferred to the west of Banda Neira (McCaffrey et al., 1984) that, with a significant component of oblique slip, may produce a large tsunami if the rupture zone parameters are large enough. It is also possible that an earthquake-induced landslide generated the tsunami (Brune et al., 2010). Nevertheless, the extent of which the earthquake was felt, and the prolonged seismicity following the event, indicates that the earthquake was larger than any event occurring over the past 100 years of instrumental records.

2.1 Testing for Mega-Thrust Earthquakes

Records of geophysical events throughout the Indonesian region date back to 1600 and are mainly compiled in “The Earthquakes of the Indian Archipelago” by Authur Wichmann (1918). This volume documents several large earthquakes and tsunami throughout the Banda Arc region that are characteristic of mega-thrust earthquake events.

It is possible that one of these events happened on 1 August 1629 in the eastern Banda Sea region:

“A half hour after the termination of a violent seismic shock there formed in the sound … a high mountain of water. The tidal wave rolled westward straight against fort Nassau on Banda Neira, as well as the village on the beach …where it achieved a height of 9 fathoms [15.3m] above the springtide stand. The mole built of stone before the fort was beaten away and the water penetrated into the fort with such force, that a 3500 pound heavy mass of iron was displaced by 36 feet [11.3m].”
The records indicate that the tsunami arrived at the Banda islands 30 minutes after the earthquake occurred. Based on the tsunami propagation equation:

\[ V = \sqrt{gD} = \frac{d}{t} \]

where \( V \) is velocity of the tsunami wave, \( g \) is gravity, \( D \) is the depth of the bathymetry, \( d \) is distance between rupture and the Banda islands and \( t \) is the tsunami arrival time, the reasonable distance between the source of the tsunami and the Banda islands is about 200-300 km.

In the Wichmann records, the earthquake was also felt in Ambon, which is 210 km to the NW, but there is no record of a tsunami there. An unusually high rate of seismicity in the region followed the 1629 event and increasingly diminished after 9 years. Fifteen years after the 1629 earthquake there were several other large earthquakes in and around Ambon, some of which caused tsunami. We interpret this temporal and spatial clustering of seismicity as stress contagion from the 1629 earthquake, which caused a series of aftershocks. The record of the possible aftershocks and the distance over which the earthquake was felt argues for a large, plate boundary earthquake as the most likely source of the tsunami. The extent over which the earthquake was felt may have been much larger than noted due to the lack of European outposts keeping records in the region during the 17th century.

The only plate boundary source region of a possible mega-thrust earthquake that is within 200-300 km from Banda Neira is the eastern-most Banda arc-continent collision zone between Seram and Tanimbar (Figure 2). This section of the collision zone is segmented around the Aru Trough region (Schluter and Fritsh, 1987). We use this segment boundary to constrain the southern limit of a Seram Trough rupture and the northern limit of Tanimbar Trough rupture (Figure 2). Other plate boundary segments in the region, such as the Timor Trough, the Wetar or
Flores backarc thrust, or the Sulawesi, Sangihe, Halmahera, Cotabato, New Guinea and Philippines Trenches are too far away to generate a 15 m tsunami that arrives 30 min after violent shaking in the Banda islands (Figure 1). Seismological evidence is lacking for how the Seram, Tanimbar or Timor Troughs are segmented.

3. Method

The numerical solutions used for the model simulations are those of Satake (2002); Lee and Ma (1997) and Ma et al. 1991. The bathymetry for the model is based on ETOPO-1 data (Amante and Eakins 2009) with a grid size of 1 minute (about 1.6 km) (Figure 2). At this resolution the Banda islands of Banda Neira and Lonthor are identified, as is the shallow embayment protecting the city of Ambon (Figure 2).

3.1 Vertical sea-floor deformation

Vertical seafloor deformation of the simulated earthquakes is modeled using the method by Okada (1985), which computes ground deformation caused by faulting in a homogeneous half space. Since the fault parameters of the 1629 earthquake are unknown, we use those from focal mechanisms of instrumented, but smaller events along what is likely the collisional interface (Harvard CMT database, http://www.globalcmt.org/CMTsearch.html). Fault width and length are calculated from empirical equations (Table 1) derived from earthquake observations (Wells and Coppersmith 1994; Lee and Ma 1997).

3.2 Tsunami propagation simulation

From the vertical surface deformation and bathymetry models, tsunami propagation is computed using a finite difference method. Here we use the linear relationship between tsunami
amplitude and fault slip (Lee and Ma, 1997; Satake 1995). We assume that the water surface is uplifted instantaneously exactly in the same way as the bottom deformation. Computations were made for a total duration of 8 hours at time increments of 5 seconds. A linear model is used to compute wave amplitudes along the coastal regions of Banda Island and Ambon that are based on the ratio of computed tsunami run-up heights to observed heights, which is known as amplification factor.

3.3 Tsunami run-up amplification factor

We calibrated the amplification factor by modeling the nearby 1992 tsunami of Flores Island (Hidayat et al. 1995) and 2006 slow earthquake tsunami in Java (Koshimura 2006) (Figure 1). These two very different tsunamogenic earthquakes in the region provide the widest possible range of direct measurements of wave and run-up heights.

3.3.1. December 12, 1992 Flore Island, Indonesia, earthquake

On December 12, 1992 a Mw 7.7 earthquake occurred near the north-shore of Flores Island along the Flores backarc thrust system (Figure 1). The earthquake and the ensuing tsunami killed around 4000 people. Most tsunami run-up heights along the northern shore of Flores Island were 2 to 5 m. Here we reproduce the same fault model used by Hidayat et al. (1995) who derive fault parameters from inverting teleseismic, broadband P and SH waves, as well as PP waves for the seismic moment rate tensor. The model uses a two-fault source to generate vertical sea-floor deformation. The fault parameters and vertical deformation can be found in Hidayat et al. (1995). The only difference between earlier models and ours is the higher bathymetric resolution we use. Hidayat et al. (1995) use the ETOPO-5 data with grid size of 5 minutes, versus the much higher resolution ETOPO-1 model we use with a 1 minute grid size. Previous
studies (Satake 1995) show that observed and theoretical tsunami heights are closer with a smaller grid size.

Tsunami run-up heights were measured along the northeast coast of Flores Island by multiple international survey teams (Table 2). However, only one tide gauge, Palpo, which is 650 km north of Flore Island, recorded open ocean wave heights (Figure 3). Comparing wave heights computed by our model waveforms with those observed by the tide gauge yields an amplification factor of 1.5 for this tsunami (Table 2). These results improved upon those of Hidayat et al. (1995) by better fitting the observed run-up heights. Locally, tsunami run-up heights were amplified by a submarine landslide caused by the earthquake.

3.3.2. July 17, 2006 South off Java Island, Indonesia, earthquake

A large earthquake (Mw 7.7) occurred along the subduction interface south of Java Island on July 17, 2006 that generated a tsunami causing around 800 casualties. Maximum run-up heights along the southern shore of Java are from 1 to 4.6 m (Hariri and Bilek, 2011). Here we reproduce the fault model in Koshimura (2006) based on Harvard CMT fault parameters, but with the higher resolution bathymetry (ETOPO-1 versus ETOPO-2). Run-up heights (Table 3) are taken from those measured along the southern Java coast by Kongko et al. (2006). Comparison of our modeled tsunami heights with the observed data indicates an amplification factor of around 4 in this case.

However, the Java earthquake is a unique case of slow rupture that was hardly felt by those on shore (Koshimura, 2006). Kato et al. (2007) made GPS observations and tsunami heights measurements during the event period, but no co-seismic displacement was detected. In their paper, the observed tsunami heights are systematically higher than those predicted from
numerical simulations based on seismic wave analysis, which indicates that fault offset may be larger than estimates using seismic analysis and the rupture was very slow. Thus, in our model we set the maximum limit of amplification as a factor of 4.

In summary, the tsunami run-up amplification factors of 1.5 and 4 that we determined from modeling two different earthquakes in the eastern Indonesia region provide upper and lower limits on the moment magnitudes for earthquakes we model on the Tanimbar and Seram Troughs.

4. Results

The most likely active faults that could produce the violent shaking (felt over 300 km radius), a tsunami with run-up heights at least up to 15 m, 9 years of possible aftershocks and perhaps stress contagion to nearby faults are the Tanimbar (South Source) and Seram (East source) sections of the Banda arc-continent collision zone. The Seram Trough is from 220-300 km to the east of Banda Neira, while the Tanimbar Trough is at least 400-440 km away, respectively (Figure 2). We primarily use the tsunami arrival times and wave heights predicted by earthquakes of various magnitudes to constrain which of the source regions is most likely. The direction of the wave and the fact that it was not observed in Ambon helps us further constrain the most likely source region. We follow our modeling procedure to identify which fault parameters and range of earthquake magnitudes for these two potential sources best coincide with the historic records.

4.1 Tanimbar Trough (South source)

For fault parameters of the Tanimbar Trough we extrapolate focal mechanism data associated with the 2004/10/22 earthquake event recorded in the Harvard CMT database (Tables
1 and 4), which is consistent with seismic reflection profiles of the plate boundary interface (Schluter and Fritsch, 1987). The epicenter of the earthquake is at 7.27° S and 130.5° E. The hypocenter is on the plate boundary interface shown by Welc and Lay (1987). A simple two-fault segment model is used to match the curved shape of Tanimbar Trough.

A vertical crustal deformation of 6 meters is required to produce a tsunami as large as that observed in the Banda Islands with the maximum amplification factor of 4 (Figure 4). According to the scaling relations, this deformation would require an earthquake of Mw 9.2 as a minimum magnitude. However, the maximum earthquake magnitude with amplification factor of only 1.5 is 9.8 for a tsunami run-up of 15 meters in Banda Islands. The finite-element-based tsunami simulation predicts a tsunami arrival time at the Banda Islands of about 58 minutes after the initiation of the earthquake (Figure 5). Although the numerical model predicts tsunami heights of 15 m, which are close to those observed at the Banda Islands, it also predicts first arrival tsunami heights > 5 m for Ambon, which was not observed.

4.2 Seram Trough (East source)

For fault parameters of the Seram Trough we use focal mechanism data from the 1993/12/04 earthquake (Table 4), which is along the boundary interface. A simple two-fault segment model is employed to match the curved shape of the Seram Trough fault zone with L=500 km, W=97 km and D=116 cm (Figure 6 and Table 4). The epicenter of the model event is 4° S and 131.14° E. The maximum earthquake magnitude of Mw 8.8 is required to generate a 15 m tsunami in the Banda Islands if the minimum amplification factor of 1.5 is applied. The minimum earthquake magnitude of Mw 8.2 is estimated with a tsunami amplification factor of 4. The Seram trough source simulation yields a Slip = 11.6 m in order to produce a 15.3 m tsunami.
This result is consistent with ‘Plafker’s rule of thumb’ (Okal and Synolakis, 2004) that a seismic dislocation does not produce run-up heights much in excess of its own amplitude of slip. The minimum estimate of tsunami arrival at the Banda Islands is 30 minutes after the initiation of the earthquake (Figures 7 and 8). Another important result is that the best-fit model tsunami produces wave heights in Ambon of less than one meter, which is consistent with the observations of feeling the earthquake, but no tsunami noted.

4.3 Sensitivity Analysis

We conducted a sensitivity analysis of our numerical methods in order to better understand how variations in hypocentral depth and fault dip from a Seram Trough earthquake influence modeled tsunami heights (Figure 9). The other fault parameters (Table 4) remained fixed for each model. For analyzing the influence of fault dip we vary dip values by increments of 5° between dips of 10° and 45°. For analyzing the influence of hypocentral depth we vary depths by increments of 5 km from depths of 10 to 30 km.

Changes in the dip angle of the fault influence tsunami height most, by a factor of > 2 between 10° to 45° (Figure 9a). This is not surprising due to the increase in vertical displacement caused by increasing dip angles. Variations in hypocentral depth have very little influence (< 10 %) on tsunami height.

5. Discussion

5.1 Evaluation of the source of 1629 Banda Mega-thrust earthquake and tsunami

The poorest fit between model predictions and observation are with the Tanimbar Trough source. There are four important misfits: 1) waves approach Banda Neira from the south rather
than from the east as observed in the record, 2) a much larger magnitude is required along the Tanimbar Trough (Mw = 9.8 to 9.2) to produce the observed run-up heights in Banda Neira, 3) the wave takes twice as long to arrive as observed, and 4) run-up heights in Ambon are at least 5 m, which would have destroyed much of Ambon. Therefore, we conclude that the most likely source of the 1629 mega-thrust earthquake is the Seram Trough.

We are also confident that the fault parameters we estimated for the Seram Trough are viable and not influencing the model adversely. High tsunami run-ups in Banda Neira may also result from local bathymetric effects that are below the resolution of available bathymetric data. However, this uncertainty does not change the better fit of a Seram Trough versus a Tanimbar Trough source, or the requirement for an earthquake along the Seram Trough larger than any observed since 1629.

5.2 Implication to tsunami hazard for Eastern Indonesia

Only one earthquake > Mw 8.0 is recorded instrumentally from the eastern Indonesia region. It happened in 1938 in the middle of the Banda Sea north of Tanimbar. Little is known about the event (Okal and Reymond 2003), but the tsunami was small and caused no damage. In 1899 a Ms = 7.8 earthquake struck the Seram region and generated a 10 m tsunami, that was perhaps landslide-assisted, killing around 3500 persons (Brune et al., 2010). With elastic strain energy accumulating across the Seram Trough at rate of 50-80 mm/a (Bock et al. 2003), it is likely that earthquake recurrence intervals are short (~100 years).

With little to no large earthquakes along the Seram Trough since 1899 and perhaps much earlier, we are concerned about the hazard potential of this very active tectonic region. During the past century population and urbanization in the region has increased ten-fold, particularly in
coastal inundation zones (Figure 1). The tsunami potential also threatens large cities outside the region. For example, our Seram Trough tsunami model predicts a 3 meter wave in Dili, which is the capital of Timor Leste and a 5.3 meter wave in north coast of Tanimbar (Figure 10 and Table 5).

6. Conclusion

This study reveals for the first time the potential of the Seram Trough for generating subduction interface mega-thrust earthquakes and associated large tsunami that carry the destructive force of the earthquake to many urbanized coastal areas of the eastern Indonesian region. The Banda Sea region has the highest number of historic tsunami in Indonesia (Hamzah et al. 2000). Yet, it also is one of the regions least aware of tsunami hazards. We urge local officials and those in harms way to not underestimate the disaster potential of the active Banda arc-continent collision, and imminent threat of tsunami hazards.
7. References


8. Figures and Captions

Figure 1 | Map of major cities and proximity to major plate boundary segments in the Indonesia region. The Indo-Australian Plate moves NNW at a rate of around 70 mm/yr. The Pacific Plate moves WNW and side-swipes the northern part of the Banda Arc at a rate of 110 mm/yr. Blue stars - epicentral locations and magnitudes of earthquakes mentioned in the paper.
Figure 2| Relief and bathymetry model of the Banda Sea and surrounding islands of the region. Active faults are shown in orange. The bathymetry is based on ETOPO-1 data with a grid size of 1 minute. At this resolution the Banda islands of Banda Neira and Lonthor (inset) are identified, as is the shallow embayment protecting the city of Ambon. Two fault plane solutions we used for the modeling are marked on this map. See section 4.1 and 4.2.
Figure 3: Comparison of observed tsunami waveform (blue) and computed waveform (red) in the tide gauge, Palpo, which is 650 km north of the December 12, 1992 Flore Island, Indonesia, earthquake event.
Figure 4: The vertical component of surface deformation estimated from the two-fault-segment model of the Tanimbar forearc. The fault plane parameters can be found in Table 4. Uplift is shown as red and subsidence in blue. The city of Ambon and the simulated gauge station are indicated as a red triangle. Triangle 1 and 2 represent the simulated gauge stations for Banda Neira and Lonthor islands.
Figure 5] Computed tide gauge records and tsunami waveform for Banda Neira and Ambon City stations from the Tanimbar Trough model earthquake. The tsunami arrival time at the Banda Islands is 58 minutes after the initiation of the earthquake, which is nearly twice as long as observed. The model also predicts tsunami run-up heights > 5 m at Ambon City using the minimum amplification factor of 1.5, which was not observed.
Figure 6] The vertical component of surface deformation for the Seram forearc, estimated from the two-fault-segment model. The fault plane parameters are given in Table 4. The color scheme and stations are the same as Figure 4.
**Figure 7** Snapshots of the first 60 minutes of tsunami propagation for the two-fault-segment model of a Seram Trough earthquake. Positive wave amplitude is shown in red and negative wave amplitude is shown in blue. The large positive wave amplitude hit the Banda Islands after 20 minutes and remains large in shallow coastal areas.

**Figure 8** Computed tide gauge records and tsunami waveform for Banda Neira and Ambon City stations from the Seram Trough model earthquake. The best-fit model tsunami produces wave heights in Ambon of less than one meter, which is consistent with no mention of the tsunami there. The tsunami arrival time at the Banda Islands is 30 minutes after the initiation of the earthquake.
Figure 9 | Sensitivity of tsunami amplitude to fault parameters for Seram Trough. Amplitude as function of fault dip (a) and hypocentral depth (b).
Figure 10 | Minimum tsunami run-up estimates for coastal cities from 1629 Seram Trough mega-thrust earthquake simulation. Star is the epicenter for the simulation.
9. Tables

**Table 1** | Empirical equation used for all numerical simulations, based on Wells and Coppersmith (1994)

<table>
<thead>
<tr>
<th>Empirical equation of thrust fault</th>
</tr>
</thead>
<tbody>
<tr>
<td>[ M_w = (5.00 \pm 0.22) + (1.22 \pm 0.16) \log(L) ]</td>
</tr>
<tr>
<td>[ M_w = (4.37 \pm 0.16) + (1.95 \pm 0.15) \log(W) ]</td>
</tr>
<tr>
<td>[ M_w = (6.52 \pm 0.11) + (0.44 \pm 0.26) \log(D) ]</td>
</tr>
</tbody>
</table>

**Table 2** | 1992 Flores earthquake observed and calculated tsunami run-up heights (in m) based on Hidayat et al. 1995. The tsunami run-up heights in last column of this table are predicted by our model.

<table>
<thead>
<tr>
<th>Station</th>
<th>Name</th>
<th>Lat</th>
<th>Lon</th>
<th>Observed (m)</th>
<th>2 Faults Model by Hidayat 1995</th>
<th>Model Predicted (m) with amplification factor 1.5</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Mage, Palu Is.</td>
<td>-8.3</td>
<td>121.75</td>
<td>2.8</td>
<td>1.79</td>
<td>1.24</td>
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<td>Mausanbi</td>
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<td>Awora</td>
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<td>2.90</td>
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<td>4</td>
<td>Deteh</td>
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<td>122.03</td>
<td>2.3</td>
<td>5.88</td>
<td>3.62</td>
</tr>
<tr>
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<td>Patisomba</td>
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<td>122.15</td>
<td>4</td>
<td>4.11</td>
<td>4.97</td>
</tr>
<tr>
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<td>Nangahureh</td>
<td>-8.55</td>
<td>122.17</td>
<td>1.9</td>
<td>3.44</td>
<td>3.23</td>
</tr>
<tr>
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<td>Wailiti</td>
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<td>122.18</td>
<td>2.1</td>
<td>3.43</td>
<td>3.42</td>
</tr>
<tr>
<td>8</td>
<td>Wuring</td>
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<td>122.2</td>
<td>3.2</td>
<td>3.67</td>
<td>3.29</td>
</tr>
<tr>
<td>9</td>
<td>Maumere</td>
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<td>122.23</td>
<td>3</td>
<td>2.95</td>
<td>2.82</td>
</tr>
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<td>Waioti</td>
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<td>122.27</td>
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<td>3.46</td>
<td>3.47</td>
</tr>
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<td>Geliting</td>
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<td>3.3</td>
<td>3.63</td>
<td>3.84</td>
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<td>Egon</td>
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<td>1.8</td>
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<td>1.95</td>
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<td>Wodung</td>
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<td>2.62</td>
<td>0.01</td>
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</table>
### Table 3 | 2006 Java earthquake- observed and calculated tsunami run-up height (in m)

<table>
<thead>
<tr>
<th>Station</th>
<th>Name</th>
<th>Lat</th>
<th>Lon</th>
<th>Observed (m)</th>
<th>Model Predicted (m) with amplification factor 4</th>
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<td></td>
<td></td>
<td></td>
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<tr>
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<td>Keburuhan</td>
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<td>Bunton</td>
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<td>1.5</td>
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<td>Cikembulan 2</td>
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<td>Pengandaran 1</td>
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<td>2.9</td>
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<td>108.70</td>
<td>2.7</td>
<td>1.5</td>
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### Table 4 | Fault parameters used for numerical simulations.

#### South Source- Tanimbar Trough (Maximum Mw: 9.8; Minimum Mw: 9.2)

<table>
<thead>
<tr>
<th>Length (km)</th>
<th>Width (km)</th>
<th>Strike 1 (°)</th>
<th>Strike 2 (°)</th>
<th>Dip (°)</th>
<th>Depth (km)</th>
<th>Max Slip (m)</th>
<th>Min Slip (m)</th>
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<tr>
<td>450</td>
<td>112</td>
<td>220</td>
<td>260</td>
<td>10</td>
<td>20</td>
<td>40.9</td>
<td>14.96</td>
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</table>

#### East Source- Seram Trough (Maximum Mw: 8.8; Minimum Mw: 8.2)

<table>
<thead>
<tr>
<th>Length (km)</th>
<th>Width (km)</th>
<th>Strike 1 (°)</th>
<th>Strike 2 (°)</th>
<th>Dip (°)</th>
<th>Depth (km)</th>
<th>Max Slip (m)</th>
<th>Min Slip (m)</th>
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</thead>
<tbody>
<tr>
<td>500</td>
<td>97</td>
<td>118</td>
<td>160</td>
<td>10</td>
<td>15</td>
<td>11.6</td>
<td>6.2</td>
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</table>
Table 5| Predicted tsunami run-up using minimum amplification factor of 1.5 from Mw 8.8 earthquake and maximum amplification factor of 4 from Mw 8.3 earthquake along Seram Trough.

<table>
<thead>
<tr>
<th>Major City</th>
<th>Banda</th>
<th>Ambon</th>
<th>Wahai Seram</th>
<th>Amahai</th>
<th>Tanimbar</th>
<th>Dili</th>
<th>Darwin Australia</th>
</tr>
</thead>
<tbody>
<tr>
<td>Minimum Tsunami Height (m)</td>
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<td>8.8</td>
<td>4.5</td>
<td>5.3</td>
<td>3</td>
<td>1</td>
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