Mass wasting along the NW-African continental margin

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Abstract: The NW-African continental margin is well-known for the occurrence of large-scale but infrequent submarine landslides. The aim of this manuscript is to synthesize the current knowledge on submarine mass wasting off NW-Africa with special focus on the distribution and timing of large landslides. The described area reaches from southern Senegal to Agadir Canyon. The largest landslides from South to North are the Dakar Slide, the Mauritania Slide, the Cap Blanc Slide, the Sahara Slide and the Agadir Slide. Volumes of individual slides reach several hundreds of km^3; runouts are up to 900 km. In addition, giant volcanic debris avalanches are widespread on the flanks of the Canary Islands. All headwall areas are complex with clear indications for multiple failures. The most prominent similarity between all investigated landsides is the existence of widespread glide planes, which follow the stratigraphy pointing to weak layers as most important preconditioning factor for the failures. Landslides with volumes larger than 100 m^3 are close to being evenly distributed over time contradicting previous suggestions that landslides off NW-Africa occur at periods of low or rising sea level. The risk associated with the landslides off NW-Africa, however, is relatively low due to their long recurrence rates.

The NW-African continental margin is home to a wide range of mass wasting processes, including landslides on open continental slopes over various scales, volcanic flank collapses, pyroclastic flows, canyon wall failures, and major turbidite systems. All these events occurred on a passive continental margin that is characterized by low-level intraplate seismicity. In contrast to many other passive margins, only few places yield clear evidence for gas hydrate occurrence. The margin is therefore an ideal setting to learn about submarine landslide processes in a setting with long sediment residence times and where gas hydrate dynamics are absent or at least not dominant.

The main objective of this manuscript is to present an update of the currently available database and state of knowledge regarding the distribution, the causes, the timing and the hazard potential of large landslides off NW-Africa. The new compilation of data allows to identify many similarities of giant landslides at passive continental margins, while at the same time revealing striking differences between individual events.

The NW-African continental margin

The NW-African continental margin is a passive margin that formed during North Atlantic rifting in the Late Jurassic. The modern morphology shows typical features of passive continental margins, including canyons, channels, deep-water reefs, and landslides at various scales (Fig. 1). The continental slope is characterized by slope gradients between 1° and 4° and extends down to ~2500 m water depth. The slope passes into the continental rise with gentle slopes of <1°. Several abyssal plain basins in water depths of more than 4500 m act as sediment traps for turbidity currents. These abyssal plain basins include the Cape Verde Basin, the Madeira Abyssal Plain (both slightly west of the map area shown in Fig. 1), the Agadir Basin, and the Seine Abyssal Plain (Weaver et al. 2000). A peculiarity of the NW-African continental margin is the occurrence of volcanic island groups (Canary Islands, Cape Verde Islands, Madeira Islands) and several volcanic seamounts (Sahara Seamounts and several seamounts north of the Canary Islands, Fig. 1). While the Canary and Cape Verde Islands show ongoing volcanic activity, the seamounts are old features (Ancochea et al. 1994; Guillou et al. 1996; Geldmacher et al. 2001).

The margin is characterized by very low sediment supply by rivers (Weaver et al. 2000). Upwelling along the entire NW-African continental margin mainly affects the middle-outer shelf and upper continental slope down to a depth of 500 m (Sarnthein et al. 1982, van Camp et al. 1991). Upwelling is generated by a complex interaction between the Trade Winds parallel to the coast and the Canary Current. In addition, the margin is characterized by significant dust import from the Sahara desert (Sarnthein & Koopmann 1980; Holz et al. 2004). Sedimentation rates in most areas are around 5 cm/kyr but may be much higher in specific regions (e.g. 22 cm/kyr at ODP Hole 658c off Cap Blanc, see Fig. 1 for location; de Menocal et al. 2000).

The NW-African continental margin is characterized by a complex interplay of sediment transport patterns directed downslope and alongslope. Acoustic imaging of the margin started more than 40 years ago. These early imaging data were the basis for the first maps showing the distribution of sedimentary features along the margin (Jacobi 1976; Jacobi & Hayes 1992). Updates of these maps based on additional data were presented by Wynn et al. (2000), Weaver et al. (2000), and Krastel et al. (2006; 2012).

Submarine mass movements of different scales occur at the continental slope including several very large-scale but infrequent landslides (Weaver et al. 2000). In areas where large landslides are absent, canyons and gullies provide pathways for sediment transport from the shelf to the deep sea (Wynn et al. 2002; Antobreh & Krastel 2006; Frenz et al. 2009). In addition, widespread debris avalanches are found on the flanks of the Canary (Krastel et al. 2001; Masson et al. 2002) and Cape Verde Archipelagos (Masson et al. 2008).

The most recent compilation of landslide features off NW-Africa is given by Krastel et al. (2012). They describe four mega-slides in an open slope environment in an area between 12° and 26°N, namely the Dakar Slide, the Mauritania Slide, the Cap Blanc Slide, and the Sahara Slide from south to north (Fig. 1). Each of these slide affected over 30,000 km^2 of seafloor. The headwall areas of the landslides show a complex morphology suggesting multiple failures and upslope retrogression. Buried landslides imaged by seismic data suggest a long history of mass wasting. Several authors suggest that all major slides in the last 200 kyr occurred during sea-level lowstands or periods of sea-level rise (Georgiopoulou et al. 2010; Krastel et al. 2012). New data have been collected in the past years and allow to extend the map presented by Krastel et al. (2012) to an area north of the Canary Islands to the Agadir Canyon region (Fig. 1). New data are also available for the head region of the Sahara Slide region. These data have important implications for the timing of landslides in this area, and are presented in this manuscript together with a summary of previous work.

Database

Sediment echo sounder, multibeam bathymetry, and high-resolution airgun seismic data of mass transport deposits along the NW-African continental margin have been collected during several cruises. These data were used as basis for selecting coring stations. Cores were usually taken by a gravity corer. Coring targeted landslide deposits but also the undisturbed sedimentary successions above or next to the headwalls.

All available information is combined to create a map showing the landslide distribution off NW-Africa (Fig. 1). Areas not covered by new data were adopted from previously published maps (e.g., Wynn et al. 2000; Weaver et al. 2000).

Major submarine landslides off NW-Africa

The term submarine landslide is used in the following as a general denomination for all mass movements generated by submarine slope failures, i.e. translational slides, slumps, and debris flows. The term glide plane is used if the base of the landslide is characterized by a widespread prominent seismic reflector following stratigraphy.

The main landslides at the NW-African continental margin in an area between 12°N und 32°N are the Dakar Slide, the Mauritania Slide, the Cap Blanc Slide, the Sahara Slide, and the Agadir Slide (Fig. 1). In addition, abundant landslides are found on the flanks of the Canary Islands. Each of these landslides will be described in the following sections.

Dakar Slide

The Dakar Slide is a giant landslide off Senegal and The Gambia. Data have been collected during RV Meteor-Cruise M65/2 in 2005 and RV Merian Cruise MSM11/2 in 2009 but data coverage is still relatively sparse (Fig. 2). The most detailed description of this landslide is given by Meyer et al. (2012). The Dakar Slide affected an about 100 km wide area bordered by two large canyon systems, the Diola Canyon in the South and the Dakar Canyon in the north. The imaged part of the headwall is found in water depths between 2000 m and 3100 m. Headwall heights are up to 140 m. Headwalls typically show a stepped pattern pointing to multiple failures (inset in Fig. 2a). A prominent sidewall is imaged at the northern edge of the landslide. It runs for about 90 km in an east-west direction and reaches Dakar Canyon at a water depth of about 3900 m. Only small parts of the southern sidewall are imaged. The minimum size of the failed area is ~8000 km^2 (Meyer et al. 2012)

Dakar Canyon was partially buried by the Dakar Slide downslope of the point at which the northern sidewall meets Dakar Canyon (Fig. 2). This is well illustrated by the bathymetric map (Fig. 2b), seismic images (Fig. 2c), and a series of sediment echo sounder profiles crossing Dakar Canyon (Fig. 2d) showing incision depths of 150-300 m upslope to an almost flat relief downslope of the intersection of Dakar Slide and Dakar Canyon.

The deposits of Dakar Slide are imaged as 100 – 150 m thick transparent to chaotic seismic facies on seismic profiles (Fig. 2c); this thickness results in a minimum volume of ~1000 km^3 (Meyer at al. 2012). The deposits of Dakar slide are covered by a ~50 m thick drape. The age of Dakar Slide is difficult to estimate because it could not be cored due to the thick drape.

Meyer et al. (2012) estimated an age of ~1.2 Ma for the Dakar Slide assuming average sediment rates of 3 – 4 cm/ka for the Quaternary drape sediments.

Mauritania Slide

The Mauritania Slide is located seaward of the arcuate shaped coastline of Southern Mauritania (Fig. 1). Detailed descriptions are given by Antobreh & Krastel (2007), Henrich et al. (2008) and Förster et al. (2010). The Mauritania Slide covers an area of about 30,000 km^2. The width of the headwall is more than 100 km; the runout distance exceeds 300 km in places. The volume is difficult to estimate because not all areas are covered by high-resolution seismic data. The estimated volume is 400 – 600 km^3, the average thickness of the mass transport deposits is ~15-20 m. The headwall area is found in water depths between 600 m and 2000 m. No detailed bathymetry is available for the headwall area but seismic profiles show a stepped pattern with headwall heights between 25 and 100 m. Glide planes are traced as a strong and continuous reflector at the base of the mass transport deposits over large areas (Antobreh & Krastel (2007).

Cores show typical debrite deposits characterized by abundant clasts with internal deformation structures, such as thinning, stretching and shearing (Henrich et al. 2008). Dating of the undisturbed sediments above the debrite suggests an age of 10.5–10.9 ka for the uppermost mass transport deposit (Wien et al. 2007; Henrich et al., 2008). Förster et al. (2010) proved the existence of another slide event, which was emplaced at < 24 ka. Seismic data show older buried events, which are not dated.

Mass wasting off Cap Blanc

The continental margin off Cap Blanc is in the central upwelling zone with sedimentation rates exceeding 20 cm/kyr in places (de Menocal et al. 2000). Earlier maps showed abundant mass wasting features in this area (Jacobi 1976; Jacobi & Hayes 1992) but these could not be confirmed by Krastel et al. (2006) or any of the numerous cruises that have taken place in the area. The compilation of all available data show two areas of mass wasting: a relatively small area around the location of ODP-site 658 in about 2000 m water depth, and the very large Cap Banc Slide that, unusually, originates in about 3600 m water depth (Fig. 1).

ODP-site 658 was drilled in a major upwelling cell off Cap Blanc in a water depth of 2263 m during ODP Leg 108 (Ruddiman et al. 1988). This almost 300 m deep (below seafloor) site dates back to the Early Pliocene but a large hiatus was observed for the time interval from 0.73 to 1.573 Ma. It was generally assumed that this hiatus is a result of a sediment slide, though it was not visible on seismic data. Bathymetric data of this area show that ODP-site 658 is indeed located in a landslide scar area (Fig. 3a). The bathymetric data suggest failures along at least two glide planes, with ODP-site 658 being located in the upper scar area. The headwalls have heights of about 50 m each. High-resolution multichannel seismic data crossing the headscarp and ODP-site 658 do not show any obvious slide deposits (Fig. 3b). The landslide must have been purely erosive in this area because no mass transport deposits were identified in the ODP-site. The glide plane is remarkably continuous and is not crossing any stratigraphy. The volume of the landslide is difficult to estimate because deposits were not mapped and the head scar is not completely covered by bathymetric data. Krastel et al. (2012) estimated the lateral extent of the evacuated area to be less than 400 km^2; the missing volume is in the range of 20 km^3. Hence, this landslide is small compared to the other landslides described in this manuscript. The failure must have occurred at around 0.73 Ma based on the hiatus recovered in ODP-site 658.

The Cap Blanc slide further downslope is much larger though only parts have been mapped. The outline shown on Fig. 1 is based on sparse hydroacoustic data. A 25 m-high head scarp was found at ~3575 m water depth (Krastel et al. 2006). The approximate age of the slide is ~165 ka (Wien et al. 2007). The reconstructed areal extent of the Cap Blanc Slide exceeds 40,000 km^2. The thickness of the deposits is in the range of ~10 – 20 m resulting in very rough volume estimates of 400 – 800 km^3 for the Cap Blanc Slide.

Sahara Slide

The Sahara Slide is a mega-slide with a run-out distance of 900 km and an estimated volume of 600 km^3 (Fig. 1; Embley 1976; Masson et al. 1993; Gee et al. 1999; Georgiopoulou et al. 2010). The evolution of the slide on its way from the scar area in about ~1900 m water depth to its distal deposits in water depths of almost 5000 m is summarized by Georgiopoulou et al. (2010). Two headwalls, each about 100 m high, represent the source area of the Sahara Slide (Fig. 4a). The slide originated as slab type failure and quickly disintegrated into a debris flow. On its way down, it passed close to the Canary Islands and incorporated volcaniclastic sand from the substrate. This resulted in a two-phase debris flow consisting of a basal volcaniclastic phase overlain by a pelagic debris flow phase. The volcaniclastic debris flow layer probably acted as low friction layer, thereby explaining the unusually long run-out distance of the Sahara Slide (Gee et al. 1999; Georgiopoulou et al. 2010). The age of the Sahara slide was estimated to be ~60 ka based on cores taken from the depositional area. Georgiopoulou et al. (2009) and Krastel et al. (2012) also presented evidence for a late Holocene reactivation of the headwall. Several older mass transport deposits are imaged on seismic data, suggesting significant mass wasting in the Sahara Slide area since at least Miocene times (Georgiopoulou et al. 2007).

New data covering the headwall area (Li et al. 2016) allow a detailed investigation of the source area of the Sahara Slide. The source region of the Sahara Slide consists of two headwall areas, named upper and lower headwall area (Fig. 4a). Only the upper headwall area has a good multibeam coverage. It shows a complex pattern of scarps with a composite height of about 100 m. Some areas of the scar show a relatively smooth sea floor while blocks are visible in other parts. The blocks have diameters of up to 200 m and are 10s of meters high. The combination of bathymetric (Fig. 4a) and sediment echo sounder data (Fig. 5a) allowed to identify three levels of glide planes. The sediment echo sounder data show a relatively thin (<10 m) cover of slide deposits on top of the glide planes in close proximity to the headwall (Fig. 5a). Glide planes do not cross stratigraphy and can be traced for large distances.

The lower headwall area is not covered by multibeam data. A single multibeam line shows a scarp more than 100 m high at a water depth of ~2650 m (Fig. 4a). A sediment echo sounder profile crossing this area parallel to the slope shows some remnant blocks close to the main sidewall suggesting a similar complex morphology of the lower headwall area (Fig. 5b). Slide deposits beneath the lower headwall are much thicker and sediment echo sounder data do not penetrate and image the base of the slide.

Some cores were taken directly beneath the upper headwall on the different levels of glide planes (Fig. 4). All cores show a similar pattern of typical debrite deposits covered by a very thin (<5 cm) Holocene drape. Only Core 07 has a drape thick enough for dating. A sample was taken at 3 cm bsf (about 1 cm above the top of the slide deposits). AMS ^{14}C-dataing on monospecific samples of the planktonic foraminifera *Globigerinoides ruber* (w) gave a calendar age of only 1840 ±23 BP (Fig. 4b). Based on this date and estimated sedimentation rates, Li et al. (2016) calculated an age of 2.2 ka for the slide deposits. No other drape of any other core could be dated but a similar thickness of the drape above the slide deposits suggest a similar young age for the entire failure of the upper headwall region. This event is recorded in one piston core further downslope, south of the Canary Islands, as a linked turbidite-debrite and as a mud turbidite capping several cores in the Madeira Abyssal Plain (Georgiopoulou et al. 2009; Frenz et al. 2009).

Agadir Slide

The Agadir Slide was already shown on previous maps (Weaver et al., 2000; Wynn et al., 2000) but has not been described in detail before. Krastel et al. (2016) presented a more detailed description based on a preliminary analysis of recently acquired data. In the meantime, a more detailed analysis of this data set has been carried out.

The Agadir Canyon region is considered as one of the most important sources for the Moroccan Turbidite System, which comprises three inter-connected deep-water basins: Agadir

Basin, Seine Abyssal Plain and Madeira Abyssal Plain (Fig. 1; Wynn et al. 2002). The Moroccan Turbidite System hosts a long sequence of turbidites, mostly sourced from the Agadir Canyon region and the Canary Islands. (Frenz at al. 2009; Hunt et al. 2013a). Individual flows generated turbidites with a sediment volume of > 100 km^3, although these large-scale events are relatively infrequent with a recurrence interval of ~10,000 years over the last 200,000 years (Wynn et al. 2002).

The headwall area of the Agadir Slide is located about 200 km south of the head region of Agadir Canyon (Fig. 6). The upper headwall cuts back to about 500 m water depth. Headwall heights are up to 100 m. The upper section of the evacuated area is narrow (only ~2 km) but it quickly widens to about 10 km at ~1300 m water depth. The upper headwall area is located in an area of pronounced sediment waves (Figs. 6, 7). A seismic profile crossing the upper headwall area (Figs. 7a, c) clearly images a set of sediment waves above a basal reflector. The seismic profile crosses the headwall slightly oblique, which makes it difficult to distinguish between the headwall and the seaward dipping flanks of the sediment waves. However, the location of the headwall is clearly visible on bathymetric data. In addition, the reflection patterns can be easily correlated between the individual sediment waves and tracing of reflectors proves that the near-surface sedimentary succession above the Agadir Slide headwall is 60 ms two-way travel time thicker than that below the headwall. Hence, about 60 ms of sediments (~50 m assuming a sound velocity of 1600 m/s for water-saturated sediments) have been removed during the failure. It is remarkable that the sediment waves were not destroyed by the landslide. The landslide removed an about 50 m thick sedimentary succession above a glide plane leaving the sediments wave beneath untouched.

A lower headwall area is located in about 1800 m water depth (Fig. 6). This lower headwall area looks subdued suggesting that it was modified by the sediments being mobilized during the failure in the upper headwall area. Headwall heights in this area reach up to 50 m.

Significant deposition of mass transport deposits starts about 30 km downslope of the upper headwall area in water depths of ~1600m (Fig. 6). Slope gradients are reduced to less than 0.6° in this area (Fig. 6b). The deposition of mass transport deposits is partly controlled by several morphological highs. These highs are the surface expressions of underlying salt diapirs (Fig. 7). Thick mass transport deposits are found landward of a small morphological high. The base of the slide can be hardly imaged on sediment echo sounder data because the deposits are too thick (>50m, see inset of Fig. 7a). Mass transport deposits seaward of the morphological high are significantly thinner. Average thickness is around 10 m downslope of the morphological high but increases to ~20 m seaward of an internal scarp (Fig. 7a). Two distinct glide planes are clearly visible on the sediment echo sounder data. These two levels represent the glide planes for most parts of the Agadir Slide.

Further down, the Agadir slide deposits can be traced in the so-called fairway (Fig. 6). The average slope angle along the fairway is 0.3°, with water depths ranging from 1900 m at 75 km below the main slide headwall, to 2600 m where the fairway enters Agadir Canyon, ~200 km below the main slide headwall. Cores collected in this area show a 4 to 5 m thick undisturbed sediment drape on top of typical sheared and contorted mass transport deposits (Fig. 7b). The estimated age of the slide deposits is ~145 ka (Mehringer 2016).

The landslide entered Agadir Canyon at about 2600 m water depth. The slope gradient increases to more than 1° when entering the canyon (Fig. 6). Core and seismic data show that the Agadir Slide uitilised the canyon for over 180 km. These data indicate that the Agadir Slide event did not transform in a turbidity current upon entering the canyon, despite slope angles increase to >1° at the fairway termination. Hence, it is unlikely that the Agadir Slide significantly contributed to the fill of the Moroccan Turbidite System (Krastel et al. 2016).

The total volume of the evacuated area is estimated to be ~170 km^3. Main evacuation occurred in the headwall areas and in the central fairway. The volume of the deposits sums up to ~340 km^3. The majority of the deposits is found in the fairway but ~70 km^3 of mobilized sediments are deposited in the Agadir Canyon. The landslide was highly erosive on its way down the continental margin.

Submarine landslides around the Canary Islands

The Canary Islands are a large volcanic islands group located at the NW-African continental margin (Fig. 1). Landslide activity on the flanks of the Canary Islands was studied in detail (Krastel et al. 2001; Masson et al. 2002). A detailed review is beyond the scope of this manuscript and here we give only a short summary of the mass wasting history. Numerous submarine landslides (mainly debris avalanches) are found on the flanks of the Canary Islands. The debris avalanche deposits can be usually linked to a well-defined amphitheater at their head, which marks the location of the corresponding flank collapse. Some debris avalanches evolve into debris flows (e.g. the Canary debris flow, Fig. 1) or can also initiate turbidity currents that are capable of flowing considerable distances downslope (Wynn & Masson 2003; Masson et al. 2006; Hunt et al. 2013a).

At least 12 giant submarine slides were identified on the flanks of the Canary Islands based on morphological data (Krastel et al., 2001; Masson et al., 2002). These slides are estimated to be younger than 1.5 Mio years because they are not buried under a thick sedimentary cover. It is interesting to note that major landslides were identified on the flanks of all islands except for La Gomera (Fig. 1). In contrast to other volcanic island groups, volcanic activity is not focused in the Canary Archipelago and the individual islands are characterized by a long history of volcanic activity. This long history of activity is balanced by a long history of mass wasting (Krastel et al., 2001) resulting in a concentration of young submarine slides around the younger western islands but also in significant mass wasting on the flanks of the old but still active eastern islands. The recurrence rate of major slides is in the range of 100 -150 ka.

Discussion

Distribution and style of mass wasting along the NW-African continental margin

The NW-African continental margin hosts numerous submarine mass movements of different types and scales. Several giant landslides, which are among the world's largest, present the most remarkable features along the entire margin from Senegal in the south to Morocco in the north. Small landslides are difficult to detect without full bathymetric coverage and may exists along the NW-African continental margin. However, numerous hydroacoustic datasets have been collected during transits between the major landslide areas. These data do not show indications for abundant small landslides and therefore we suggest that small landslides are indeed an exception along the NW-African continental margin. Power-law and/or lognormal relationships for the cumulative volume distribution of submarine slope failures have been suggested for some areas (ten Brink et al., 2006; Chaytor et al., 2009; Urgeles and Camerlenghi, 2013). We are aware that any statistical analysis of the landslides presented in this manuscript does not make sense due to the small numbers of landslides but we want to point out that abundant smaller events along the NW-African continental margin seem to be absent; there is a clear imbalance towards very large landslides.

The distribution of landslides off NW-Africa does not show any obvious regularities. Some are located in arcuate margin sections (Mauritania Slide, Dakar Slide); the Cap Blanc Slide is located offshore a major cape and others are located at relatively straight sections of the margin (Sahara Slide, Agadir Slide). Most of the landslides are bound by major canyons (Fig. 1). These canyons/channels and associated levees appear to present a barrier for the landslide. Only the Dakar Slide destroyed the lower reaches of the Dakar Canyon at its northern boundary. Another example for interaction is the Agadir Slide, which entered Agadir Canyon and continued to move as a confined flow for over 180 km inside the canyon.

Krastel et al. (2012) suggested that large submarine slides are preferably found in regions where major canyons are absent between the bounding canyons. Canyons along the NW-African continental margin are well known as important pathways for downslope sediment transport (Hanebuth & Henrich 2009; Henrich et al. 2010; Pierau et al. 2010), while the areas without canyons become increasingly burdened by deposition of thick sedimentary successions. These thick sedimentary deposits may be especially prone to failure. However, the area with the highest accumulation rate in the investigated area is found off Cap Blanc (Martinez et al. 1999, de Menocal et al. 2000). No major canyons are found in this area (Fig.

1). The giant Cap Blanc slide is found off Cap Blanc but it originated at the continental rise at water depths of ~3575 m. Accumulation rates in this depth are much lower than at the upper and middle continental slope. The landslide around ODP-site 658 at the middle continental slope is only small leaving a large area with a thick unfailed sedimentary succession off Cap Blanc. Hence, sediment transport through canyons may prevent the accumulation of widespread thick deposits but the absence of canyons does not imply that such areas are especially prone to fail.

The landslides along the NW-African continental margin show several similarities but also some differences. Most of the landslides originate in water depths between ~500 m and 2000 m, with the exception of the Cap Blanc slide with the main head scarp in the unusual large water depth of ~3575 m. All headwall areas are complex with clear indications for multiple failures most likely occurring in retrogressive style.

The combined interpretation of the headwall morphologies and cores suggest that most landslides originated as slab-type failure. Failure surfaces commonly appear very smooth and are parallel to sediment bedding at different stratigraphic depths. Disintegration and transformation into debris flows is typical for most slides off NW-Africa. Most of the flows seem to be highly erosive resulting in much larger volumes of the mass transport deposits compared to the evacuated volumes in the landslide scars. Some transformed into highly mobile turbidity currents soon after slope failure (Wynn et al., 2002), whereas others did not record such a transformation. For example, the Sahara Slide offshore Western Sahara has a run-out distance of up to 900 km, it failed as a translational slide and disintegrated into a plastic debris flow. However, there is no evidence for its transformation into a turbidity current (Gee et al., 1999; Georgiopoulou et al., 2010). Yet, all landslides have essentially similar compositions and an explanation for such fundamentally different flow behaviours is not well understood.

Preconditioning factors and triggers

Numerous processes that may act as preconditioning or triggers for submarine landslides have been suggested in the past. However, in almost all cases it is difficult to relate individual landslides to specific processes, and this is also the case for landslides off NW-Africa. Preconditioning factors that may have played a role for landslide generation off NW-Africa include (1) excess pore pressure due to sediment deposition, (2) the presence of gas, and (3) the occurrence of weak layers (Canals et al. 2004). Owing to the morphological similarity of large landslides off NW-Africa, we assume that similar preconditioning factors apply.

It is well accepted that high excess pore pressure can be generated due to rapid sediment deposition (Binh et al. 2009; Stigall & Dugan 2010). However, Urlaub et al. (2015) showed that sedimentation rates as observed off NW-Africa (<20 cm/kyr) are usually not high enough in order to generate sufficient excess pore pressures to directly cause failure of a 2-4° slope, which are typical slope gradients in the headwall areas of the landslides along the NW-African continental margin. Hence, other factors need to be considered. The presence of gas in the sediment may lead to excess pore pressure if unloaded or a continuous source exists. Unloading may relate to erosion or sea level fall. The dissociation of gas hydrates or focused fluid flow through fluid pathways may form gas sources that lead to excess pore pressure if trapped in the sediment column. We do not see any indications for gas hydrates in our data but the sediments off NW-Africa do have a relatively high gas content due to burial of high amounts of biogenic material. In ODP-site 658, the gas concentration drastically increases at sub-bottom depth greater than 25 m (Mienert & Schultheiss) but only a relatively small landslide is observed in this area. Antobreh and Krastel (2007) report fluid escape features for the Mauritania Slide and suggest excess pore pressures as an important trigger mechanism for the formation of the Mauritania Slide. However, fluid escape features are not evident for other landslides off NW-Africa. Hence, focused fluid flow may have played a role for some slides but it is unlikely that this is the main preconditioning factor. The most prominent similarity between all landsides found along the NW-African continental margin is the existence of a strong continuous reflector at the base of the slide, which follows the stratigraphy. We are aware that the resolution of the seismic data is limited but we do not see any irregularities at the base of the slides on datasets of different resolutions suggesting a bedding parallel failure. The most likely reason for such failure behavior are widespread weak layers (Masson et al. 2010). While the weak sections would have been removed during the failure, the imaged glide planes must represent the stronger and more resistant sediments immediately beneath the weak layer. The composition of such weak layers is largely unknown because they are removed by the landslide and/or not cored. Urlaub et al. (2015), however, showed with numerical and experimental modelling that the presence of a layer of high compressibility has the potential to generate slope failures at slopes with a slope gradient of 2° and accumulation rates as observed off NW-Africa. An interesting aspect in this context is the dust input from the Sahara Desert to the continental margin. Aeolian dust input varied over time; it dominated the arid glacial periods during isotope stages 4, 3, and 2 (Zühlsdorff et al. 2007). Henrich et al. (2010) postulated climate and sea level induced turbidite activity in NW-African canyon systems. For example, voluminous turbidites frequently passed through the Timiris Canyon during sea-level lowstands. Silty dust supplied by strengthened trade winds is considered as main source of these turbidites (Henrich et al. 2010). Thick piles of dusty sediments along the margin might be essentially prone to failure but dust supply may also act as fertilizer by delivering iron and therefore enhancing primary productivity resulting in the deposition of sediments being rich in organic matter and microfossils. Such deposits may form layers of high compressibility, which have the potential to fail at low slope gradients (Urlaub et al. 2015). In summary, we consider weak layers as the most important preconditioning factor because all slides occurred along widespread surfaces parallel to the stratigraphy. High accumulation rates and high gas concentrations may also play a role but to a lesser extent. The final trigger for the landslides remains highly speculative. Despite being a passive margin, it is known that large earthquakes (M > 7) may occur within 'stable' plate interiors at long recurrence intervals (Calais et al. 2016). These earthquakes may act as trigger but several investigations showed that even the largest megathrust earthquakes are not capable of causing large, deep-seated landslides unless the slope is preconditioned in some ways (e.g., Völker et al. 2011).

Timing and hazard potential

Large submarine landslides present a major geohazard due to the capability to destroy offshore infrastructure and to trigger tsunamis. Tsunami heights, however, are extremely difficult to estimate for large submarine landslides because their failure dynamics is usually unknown. The critical factors for the tsunami generation capacity of submarine landslides are: (1) shallow-water to intermediate depths (<1,000 m), (2) significant volumes (>2 km^3), (3) stiff cohesive material (e.g., consolidated clay), and (4) rapid initial acceleration of the failed material (Watts et al. 2000; Ward 2001, Harbitz et al. 2006). Most of the landslides along the NW-African continental margin fulfill these criteria though several factors are difficult to estimate. All failures except for the Cap Blanc Slide occurred in the right water depth range for generating tsunamis. The landslides on the flanks of the Canary Islands started as subaerial landslides and entered the ocean. The volumes of the investigated landslides exceed 100 km^3. All slides show clear indications for failure in multiple stages, and the volumes of the individual failures are difficult to estimate. However, it is very likely that individual failures mobilized several 10s of km^3 of material. The slides originated as slab-type failure with a high tsunami generation potential but most of them quickly disintegrated into debris flows. The initial acceleration remains the largest unknown as no data are available for this parameter. There are only a few indications that past landslides along the NW-African continental margin triggered or did not trigger a tsunami. Henrich et al (2008) speculate that the Mauritania Slide triggered a tsunami. They observed that the debrite deposits are directly overlain by a turbidite bed containing shelf material. Hence, the main failure of the Mauritania Slide may have triggered a tsunami, which in turn mobilized shelf sediments and released the turbidity current. For the Sahara Slide, Georgiopoulou et al. (2009) observed a major turbidite in the Moroccan Turbidite System, which is interpreted to have formed during the most recent failure of the steep Sahara Slide headwall due to an almost simultaneous generation of both a debris flow and a turbidity current, or by entrainment of water and progressive dilution of the debris flow leading to the formation of an accompanying turbidity current. This process would suggest a relatively fast moving landslide body having a high tsunamigenic potential. In contrast, no turbidite is associated with the 60 ka failure of the Sahara Slide suggesting a slow moving slide with a relatively low tsunami generation capacity.

Large volcanic debris flows associated with volcanic island flank collapses may cause devastating tsunamis as they enter the ocean. Computer simulations show that the largest of these volcanic debris flows on oceanic islands such as Hawaii or the Canaries can cause ocean-wide tsunamis (e.g. Løvholt et al., 2008) though it is extremely difficult to model these tsunamis because the failure dynamics and volume is unknown. It is relatively easy to estimate the total volume of the failure deposits but recent studies show that the total landslide volume resulted from multiple failures with much smaller volumes of individual failures (Hunt et al. 2013b). However, several flank failures at the Canary Archipelago are well documented by large debris avalanche deposits and corresponding tsunami deposits on the islands (Paris et al. 2004; 2017; Pérez-Torrado et al., 2006).

In order to estimate the risk associated with failures along the NW-African continental margin and on the flanks of the Canary Islands, a critical parameter is the timing and the frequency of the failures. Failures of the flanks of the Canary Islands would have catastrophic consequences due to the failure itself and the large tsunamis, which would be triggered. The frequency of such giant failures, however, is extremely low. Krastel et al. (2001) suggest one major failure every 100 – 150 kyr for the Canary Islands. Hence, the risk associated with these failures is low due to the long recurrence periods.

The risk related to landslides along the continental margin is more difficult to assess because the timing of failure is not always well known. Lee (2009) compiled ages of landslides in the Atlantic Ocean. He observed that most dated slides occurred during times of a rising sea level. As a consequence, the Atlantic Continental margins would be relatively stable during the present time of a high, almost constant sea level. This observation is doubted by Urlaub et al (2013). They note that fewer landslides occurred in the past 6 kyr but clearly state that this pattern is not statistically relevant. For the NW-African continental margin, we find that the known ages of dated landslides with volumes larger than 100 m^3 are close to being evenly distributed (Fig. 8).

The upper headwall of the Sahara Slide was most likely generated by a failure, which happened only about 2 ka BP (Li et al. 2016). This event shows that landslides along the NW-African continental margin happened during periods of a high constant sea level. We want to point out that the young age of the Sahara Slide was only found due to detailed investigations of the headwall area. Distal deposits have been dated at 60 ka (Gee et al., 2009). This example shows that detailed investigations from the headwall area to the distal deposits are necessary in order to understand the full history of mass wasting in a specific area. Additional data in headwall areas with less dense data coverage (e.g. Mauritania Slide and Dakar Slide) may also reveal young reactivations of headwalls, which would enlarge the tsunami hazard.

To sum up, the largest hazard caused by mass wasting in the investigated area is associated with large flank collapses on the Canary Islands. Such failures, however, are very rare events. Failures along the continental margin are more frequent and have also taken place in the recent past. Nevertheless, the probability of such failures is still relatively low. The failures are definitely large enough for generating tsunamis but realistic estimates for the tsunami heights are extremely difficult due to the unknown failure dynamics.

Conclusions and outlook

The new compilation of all available data allowed to produce a map showing the distribution of submarine landslides along the NW-African continental margin from Senegal in the south to the Agadir Canyon region (Morocco) in the north. The following conclusions can be drawn based on this compilation:

- The NW-African continental margin is characterized by large (>100 km^3 volume) but infrequent events. Large-scale debris avalanches occur on the flanks of the Canary Islands. Small landslides seem to be almost absent.
- All landslides occurred as multiple failures, most likely in a retrogressive pattern. Multiple failures resulted in smaller volumes of individual failure events, which is important when estimating the tsunami generation potential.
- The most obvious similarity between all landslides is the failure along pronounced and widespread glide planes, which are parallel to the stratigraphy. Hence, widespread weak layers are considered as most important preconditioning factor though the nature of these weak layers remain unknown.
- At least one failure along the NW-African continental margin occurred in the younger past (~2ka BP). This contradicts the postulate of a stable continental margin during periods of a high constant sea level.
- In order to investigate the full history of a landslide, it is essential to investigate the entire path of a landslide from its sources to the distal deposits because otherwise events such as a young reactivation of headwalls may be missed.
- Landslides on the flanks of the Canary Islands have the highest hazard potential but such events are very rare. Landslides along the continental margin are more frequent but still have long recurrence intervals. They have the potential to trigger tsunamis

though no confident evidence for a tsunami triggered by a landslide has been identified in the geological record so far.

Several open questions call for additional research. Main targets of future work should include i) additional seismic and hydroacoustic surveys in order to characterize slide deposits and glide planes at higher resolution, ii) targeted sampling based on drilling at various position of the slide bodies and especially outside the slide areas to understand the slope stratigraphy that is involved in the failures as the beds that act as “weak layers” are probably not preserved in the landslide evacuation area itself, and iii) sedimentological and geotechnical studies of various lithologies and lithological combinations to understand why and when weak layers are generated.

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Figure Captions:

Fig. 1: Map showing the distribution of landslides along the NW-African continental margin. See text for details.

Fig. 2: Compilation of acoustic data of Dakar Slide. (**a**) 3D perspective view of the ~100 km wide headwall area of Dakar Slide. A bathymetric profile shows a stepped pattern of the headwall. See Fig. 1 for location. Modified after Krastel et al. (2012); (**b**) Bathymetric close-up

of the intersection of a sidewall of Dakar Slide and Dakar Canyon. The morphology indicated that the canyon was destroyed by the slide (modified after Meyer et al. 2012); (**c**) Seismic profile showing the deposits of the Dakar Slide. Dakar Canyon is filled by slide deposits. Location of profile is shown on Fig. 2b (modified after Meyer at al. 2012); (**d**) Series of sediment echo sounder profiles crossing Dakar Canyon. The incision depth of the canyon is significantly reduced beneath the intersection of the Dakar Slide and the Dakar Canyon. The sediment echo sounder data show a rough and partly hummocky surface in the area of the Dakar Slide deposits. Locations of profiles are shown on Fig. 2a.

Fig. 3: (**a**) Perspective view of the scar in the region around ODP-site 658. See Fig. 1 for location of map. (**b**) Seismic profile crossing ODP-site 658 and the headwall above. A hiatus was observed for the time interval from 0.73 Ma to 1.573 Ma in ODP-site 658. This hiatus is caused by a landslide. However, no mass transport deposits are found at the depth of the hiatus pointing to a purely erosive landslide in this area. See Fig. 3a for location of profile.

Fig. 4: (**a**) Bathymetric map of the Sahara Slide scar. Two distinct headwalls named upper and lower headwall are found in water depths of 1900 m and 2600 m, respectively. The complex shape of the headwall and different levels of glide planes suggest multiple failures. A smaller landslide is found south of the main Sahara Slide. (**b**) Panel of cores collected beneath the upper headwall on different glide planes (see Fig. 4a for location of cores). A photograph (left), a schematic drawing (middle), and an interpretation is shown for each core. The Holocene drape on top of the slide deposits is < 5 cm for all cores. The only drape which could be dated was present at core 07. A sample at 3 cm bsf (1 cm above the slide deposits) was dated at 1840 ±23 BP.

Fig. 5: (**a**) Sediment echo sounder profile crossing the upper headwall area. Three distinct glide planes (GP) can be identified on this profile. Thin slide deposits are imaged as a chaotic to transparent body above the glide planes. See Fig. 4 for location of profile. (**b**) Sediment echo sounder profile crossing the lower headwall area. A smaller mass transport deposit (MTD) is imaged south of the main Sahara Slide. The edge of the Sahara Slide is characterized by mass wasting at different stratigraphic depths and a remnant block. This pattern suggests a similar complex morphology as the upper headwall area. See Fig. 1 for location of profile.

Fig. 6: (**a**) Perspective view of the Agadir Canyon and Agadir Slide (modified after Krastel et al. 2016). Agadir Slide can be traced from its headwall area ~200 km south of Agadir Canyon to the Agadir Canyon. See Fig. 1 for location of map. (**b**) Morphological profile along the path of the Agadir Side. The slope gradients of the individual sections of the slide are listed below the profile. The location of the profile is shown as yellow solid line on Fig. 6a.

Fig. 7: (**a**) Seismic profile crossing the headwall area of the Agadir Slide and the upper part of the depositional fairway (see Fig. 6a for location of profile). The inset shows a sediment echo

sounder image of parts of the profiles. Thick mass transport deposits are found upward of a small morphological high, which is caused by an underlying salt diapir. Two distinct glide planes are clearly imaged downslope of the morphological high. (**b**) Photo of Core MSM32-27. The core contains interbedded hemipelagic slope sediments down to 470 cm, underlain by sheared and contorted debrite. The location of the core is approximately 1 km northwest of the location shown on Fig. 7a. The location of the core is also marked on Fig. 6a. (**c**) Enlargement of seismic image in the upper headwall area of the Agadir Slide (modified after Krastel et al. 2016). Sediment thickness above the dashed reflector is significantly greater for the area above the headwall, suggesting removal of a substantial volume of sediment by the landslide.

Fig. 8: Histogram representation for failures off the coast of north-west Africa using an updated version of the database presented by Urlaub et al. (2013). The dark grey bars show the most likely ages. The open bars with black edges take into account uncertainty intervals, assuming an evenly distributed probability along this interval. The bin interval is 5 ka. The grey curve depicts global mean sea level (Waelbroeck et al., 2002).