Ocean Upper Mantle Initiative: The oceanic lithosphere and its interaction with mantle convection

Konzept und Planungspapier, Koordination Support Schiffsanträge Sonne, Polarstern, Meteor/Merian, Mittelgrosse Forschungsschiffe

Ansprechpartner: apl. Prof. Frank Krüger
Tel.: +49-331-977-5813, e-mail: kruegerf@geo.uni-potsdam.de
Prof. Dr. Torsten Dahm
Tel.: +49-40-42838-2980, e-mail: torsten.dahm@zmaw.de

weitere Beteiligte (die Liste ist offen)

Wolfgang Friederichs (Uni Bochum)
Wolfram Geissler (AWI)
Karsten Gohl (AWI)
Ingo Grevemeyer (IfM Geomar)
Marion Jegen (Uni Kiel, IfM Geomar)
Thomas Meier (Uni Bochum)
Robert Patzig (AWI)
Wolfgang Rabbel (Uni Kiel)
Carsten Riedel (AWI)
Georg Rümpker (Uni Frankfurt)
Karin Sigloch (LMU München)
Vera Schlindwein (AWI)
Christine Thomas (Uni Münster)
Frederik Tilmann (Cambridge, GFZ)
Xiaohui Yuan (GFZ)
High-frequency (20Hz) fin and blue whale singing recorded on DEPAS hydrophones

Low-frequency surface waves (2.09.09, M7.4 Java) recorded on DEPAS seismometer (Z)
Executive Summary

The DEPAS instrumental pool is a German large scale facility in seismology available to universities and research institutions wishing to conduct long term passive seismic experiments on the seafloor. It comprises 80 broadband ocean-bottom stations (seismometer and hydrophone with optional further instruments such as tilt) and, for onshore/offshore experiments, a dedicated pool of 100 land stations. A number of recent experiments show the great potential of this facility, which is on par with the US ocean bottom pool, and currently unique within Europe with its combination of broadband sensor technology and long deployment times (currently up to 1 year). However, this facility has not yet been fully utilised by the broadband seismological community because of the special logistical and financial requirements of ocean bottom experiments, which make it much more difficult to plan and secure funding compared to land experiments. In general, guaranteed ship time on a suitable research vessel during two cruises is required for deployment and recovery, separated by up to a year, with additional servicing cruises required for deployment times in excess of one year. Because of large consumable costs (e.g., lithium batteries), high insurance costs and the requirement for specialised technicians, the cost of, for example, a 24 instrument deployment for one year can easily reach nearly 0.25 Mill€ even before ship time is taken into account. Experiments on such scale can only be successfully funded and organised, if several groups coordinate their efforts for a common scientific goal, and if review panels are aware of the complexities but also the great promise of ocean bottom experiments.

This White Paper was developed by a group of seismologists to act as a catalyst for such cooperation by outlining the major scientific issues in the field of oceanic lithosphere-asthenosphere interaction and more broadly the convective system in the upper mantle, which until now could not be answered due to the dearth of seismic data from the oceans but could be successfully tackled by employment of the DEPAS facility. Furthermore, we would like to inform multi-disciplinary review panels, ship planning committees and other interested parties of the approach and mode of operation of ocean bottom experiments, and the current state-of-the-art in the analysis of ocean-bottom data. We discuss which techniques familiar from land seismology already work well in the oceans, and which need further development work to utilise their full potential.

Opportunities for experiments within the theme of oceanic lithosphere-asthenosphere interaction and convection exist with regard to the question of 1. how upper mantle convection causes lithospheric instabilities, and how they relate to plate formation (deviation from half-space cooling model) and anomalous provinces such as the Ontong Java plateau; 2. what is the structure and state of the upper and lower boundary of the oceanic asthenosphere, i.e. their sharpness and depth, and the sub-oceanic transition zone, in particular whether they are fundamental differences to the continents; 3. what is the state of stress within oceanic plates and at passive margins, how does this stress field relate to intra-plate earthquakes, and to what extent is it influenced by asthenospheric processes such as basal drag or edge driven convection; 4. how do triple junctions and hotspots interact with the edges of plates.

Although the seafloor represents a challenging environment for seismological experiments, in particular due to the high noise levels found on horizontal components, a number of standard seismological techniques have been applied successfully in past experiments. In particular, ocean bottom instruments are ideally suited to analyse surface wave dispersion in and below
oceanic plates (Rayleigh waves) and many experiments both within DEPAS and internationally have demonstrated that ocean bottom recordings of surface waves are at least as good as those of land seismometers. Techniques with higher lateral resolution (body wave tomography, SKS splitting, receiver functions) have been applied successfully in past ocean bottom experiments but generally at longer periods, and for stronger signals than in land experiments. However, we see good opportunities in fully harnessing these techniques by innovations in deployment designs and signal processing, for example by application of array techniques or combination of multiple sensors (tilt, pressure). Pilot type DEPAS experiments could be used to test and refine novel deployment designs and analysis methods.

In short, the DEPAS pool represents a great opportunity to address the big questions related to plate formation and mantle convection but in order to realise this opportunity, cooperation among scientific groups but also among review panels, e.g. the ship planning committees for different research vessels and the committees within the funding agencies, is essential.
2. Introduction

Fig. 1: Overview of tectonic plate boundaries of different types (red: ridges, green: transform, blue: trenches) plotted together with instrumentally recorded earthquakes of the last 100 years. Black areas indicate igneous provinces and hotspots. Yellow triangles give positions of former DEPAS experiments, and white polygons indicate regions where different groups plan future DEPAS experiments.

The concepts of seafloor spreading and global plate tectonics (Fig. 1) were introduced in the 1960s and 1970s (Vine and Matthews, 1963; McKenzie and Parker, 1967; W. J. Morgan, 1968; Isacks et al., 1968) and completely renewed our understanding of the geodynamic processes on Earth. Different processes such as the formation of oceanic crust and lithosphere at seafloor spreading ridges and the orogenesis of mountain chains could be explained by a single theoretical framework, which accounted for a large number of observations, in particular the pattern of seafloor magnetic stripes, the variation of seafloor depth and heat flow with oceanic crust age, and the concentration of earthquakes and volcanoes at plate boundaries. Plate tectonics is essentially a kinematic model describing in great detail the motions of rigid plates, whereas many of the underlying driving forces to explain the movement and behaviour of plates remained speculative at first and without direct proof because they required a detailed look into the mantle, which is challenging even today.

Even in 2010 major questions concerning the dynamic processes making plate tectonics possible are still being debated. Is plate tectonics driven by mantle convection or vice versa (top-down tectonics, see e.g. Anderson, 2001), and are driving forces different for fast and slow spreading ridges? What causes the deviation of ocean depth or seafloor topography from the predictions of a cooling half-space model? What is the balance between drag forces, ridge push and active upwelling at mid-oceanic ridges and slab pull versus frictional and drag forces close to subduction zones? How is subduction initiated? How thick is the lithosphere, and what is the physical and chemical nature of the Lehmann and Gutenberg discontinuities? Is convection stratified in the mantle? What is the role of upper mantle small-scale
convection and does it exist at all? How efficient is the lithospheric reheating associated with sub-lithospheric small-scale convection, and may it cause lithospheric delamination (e.g. Lenardic et al., 2003) and regional uplift? Is edge-driven convection an alternative to plumes for explaining hot spot volcanism (e.g. King and Ritsema, 2000)? May edge-driven convection explain the Bermuda Rise (Vogt, 1991), the North Atlantic Tertiary volcanics (e.g. Holbrook et al., 1993), and the flood basalt magmatism at the peripheries of cratonic provinces? How wet is the oceanic asthenosphere and are some hot spots really wet spots or chemo-spots? What is the role of passive plate boundaries between thicker and thinner plates: does the thicker plate cause large compressional stresses in the thinner plate, and may this generate intraplate thrust earthquakes? What is our understanding of intraplate earthquakes in general, and are they affected by asthenospheric mantle flow?

These questions are not only of fundamental importance to our understanding of how the solid Earth works but have profound consequences for other parts of the Earth system and also bear on questions of societal relevance. For example, the long term water and carbon cycles cannot be understood properly without taking into account processes at ridges and subduction zones, and significant sea level changes result from changes of average seafloor depth due to hot spot related uplift and the balance of young and old seafloor; the thermal structure of oceanic margins influences the formation of oceanic (limestone) platforms and the creation and maturation of hydrocarbon resources; lastly, intra-plate earthquakes can generate tsunamis far from subduction zones and are perhaps an underestimated hazard – widespread devastation was caused in 1755 in Lisbon by a tsunami and shaking induced by an earthquake which occurred in the Sea of Cadiz, which lacks a clearly demarcated plate boundary.

The answer to many of these questions lies in the upper mantle, and therefore it has been a focus of much seismological research to better resolve its global and regional structure. For instance, improvements in tomographic imaging methods in the last decade completely renewed our vision of the upper mantle (Fig. 2 for global S-wave velocity perturbations at 150 km depth), with clear differences between continents and oceans at depth to at least 200 km. The study of elastic anisotropy on lithospheric scale gives clues about the flow patterns in the mantle (Savage, 1999). Recent advances make it possible to simultaneously consider velocity heterogeneity and anisotropy (e.g. review by Montagner, 2007), although fundamental problems remain in separating anisotropy and velocity heterogeneity unambiguously, unless the target area is sampled by a dense array of stations. Finally, it thus became possible to tentatively relate surface geology and the motion of the plates to underlying mantle structures and convective patterns.

We briefly discuss two examples of how our view of the upper mantle has renewed during the last decade, and then discuss how ocean bottom experiments can improve on thses studies.
Gung et al. (2003), using an inversion procedure for transverse isotropy from body and surface waves, suggested that the Lehmann discontinuity, observed under continents and especially under old cratons, is a rheological boundary marking a transition from anisotropic to isotropic structure. The varying depth of the anisotropic asthenospheric channel is marked by the Lehman discontinuity under continents at a depth of about 200–250 km, and the Gutenberg discontinuity under oceans at the depth of about 60 – 80 km. The observed anisotropy in the upper mantle is flow-induced and related to present-day plate motion under both oceans and continents, but occurs at different depths.
Zhou et al. (2006) used a novel finite frequency surface wave tomography, based upon complete three-dimensional Born sensitivity kernels, and retrieved a global shear-wave velocity structure and radial anisotropy in the upper mantle down to 400 km depth. The model confirms the previous depth distribution of continental and oceanic asthenosphere and the flow pattern of mantle convection beneath old cratons and the Pacific. Distinct from previous upper mantle models is the identification of a strong negative radial anisotropy at depths greater than 120 km beneath mid-ocean ridges (see low SH velocities in Fig. 3). The depth extent of the ridge anomaly is different beneath fast- and slow-spreading centres. The Northern Mid-Atlantic ridge anomalies extend down to at least the top of the mantle transition zone at 400 km depth, whereas the South Atlantic and the fast-spreading ridges in the Pacific have no deep expression. The authors therefore suggest that active upwelling beneath slow-spreading ridges may play a major role in the opening of the seafloor there.

Imaging techniques based on converted and transmitted body waves provide a completely different way of imaging upper mantle structures on a regional scale. Well known are P- and S-wave receiver function techniques beneath seismological arrays, which are established on-shore but not yet convincingly demonstrated for ocean bottom recordings in spite of a few successes (notably the MELT experiment, Shen et al., 1998). These studies have revealed anomalies of crustal and lithospheric thickness or of the upper mantle transition zones, mostly beneath hotspots or subduction zones.
The imaging of large sections of the mantle is hampered by a lack of data from ocean bottom stations. Sampling is poorest in the southern hemisphere, where a number of mantle plumes, including two super plumes beneath the Pacific and Africans plates, likely play a large role in driving the Earth's heat engine (Simons et al., 2009). Due to the missing coverage of oceanic recordings on a regional scale, small-scale features of the oceanic upper mantle are rarely studied. On the other hand, several seafloor field data indicate that small-scale features of the oceanic lithosphere-asthenosphere system exist and possibly play an important role for plate tectonic processes. For instance, heat flow measurements and seafloor topography in the Pacific clearly deviate from simple half-space models if the plate is older than 80-90 Mio years, and indicates a possible influence of small-scale upper mantle convections (see e.g. Zhong et al., 2007; Crosby et al., 2006).

Small-scale upper mantle features can only be recovered by seismological data recorded on local- and regional-scale arrays. A good example is given by the experiments of the EarthScope/US-Array experiments, where several high density seismological broadband arrays are deployed on a local/regional scale over typical periods of years and shifted to different geographical areas over a period of several years, alongside a larger array moving across the US in a rolling manner. For instance, Yang et al. (2008) analysed fundamental Rayleigh waves recorded by such a transportable array component, collected during a ~1 year deployment. They showed that novel analysis concepts such as ambient noise tomography using surface waves at periods above 20s are well suited to resolve structural details of the upper mantle on a regional scale. Another example is given by Sigloch et al. (2008) to unravel the deep structure of the Farallon plate subduction – Yellowstone hotspot interaction at a resolution of less than ~200 km.

Surface waves play an important role in resolving the upper mantle details beneath the oceans, because they typically generate better path coverage of the upper regions of the Earth than body waves and provide a better vertical resolution. Further, they can be used to study the anisotropy of the upper oceanic mantle. Fortunately, noise conditions at the ocean floor enable Rayleigh wave surface wave studies at periods above 15s with novel wide-band ocean bottom stations (e.g. Dahm et al., 2006, Tilmann and Dahm, 2008, Webb, 1998).

Additionally, body wave studies from passive broadband deployments have superior lateral resolution in the upper mantle compared to surface waves. Several of these body wave methods are based on high fidelity recordings on horizontal components. Unfortunately, the horizontal component recordings on the ocean floor seismometers are problematic at periods longer than 10 seconds due to high noise levels, much of it tilt-induced. It is important to further improve the coupling as well as the processing techniques of wide-band ocean bottom stations. The possibility and feasibility of ocean bottom wide-band deployments, as well as perspectives of novel techniques, are further discussed in the section below.

2.1 Instrumentation
The DEPAS instrumental pool is a German large scale facility available to universities and research institutions wishing to conduct long term passive seismic experiments on the seafloor. Its broadband ocean bottom component is operated out of the Alfred-Wegener-Institute in Bremerhaven and consists of 80 stations. Each of the stations is equipped with a Guralp CMG40 T seismometer (sensitivity 60 s – 50 Hz) and HTI differential pressure gauges (sensitivity 100 s – 8 kHz). Due to their modular design, the instruments can optionally be equipped with additional instrumentation, e.g. tilt sensors. Most stations can be deployed in
water depths of up to 6000 m with deployment periods of up to 12 months, with the exception of 12 deep-sea capable instruments which can be deployed up to 7200 m water depth, but have a reduced maximum deployment time of 8 months (for details see http://www.awi.de/en/go/depas). For onshore/offshore experiments, a dedicated pool of 100 land seismometers (CMG3ESP, 60s-50 Hz) is maintained at the GeoForschungsZentrum Potsdam (Geophysical Instrument Pool Potsdam, GIPP). Scientists at German research institutions can apply for a loan of the equipment where the merit of applications received is decided by an external steering board and internal advisory board based on scientific excellence and feasibility of the proposed work. The equipment is provided free of charge, but all operational costs, including technician time and instrument insurance, need to be raised by the user. Detailed costs vary widely with the type of experiment, but as indication a 1 year deployment of 24 ocean bottom instruments can cost nearly 0.25 Mill€ even before shiptime is taken into account. Approximately two thirds of the cost relate to mobilisation/demobilisation and fixed deployment costs, and are incurred irrespective of the length of the experiment, and one third is for batteries and will scale with the length of the experiment. Additional costs, primarily for technician time, will be incurred if the duration of the experiment exceeds the maximum deployment time, and the instruments need to be recovered and redeployed halfway through the experiment.

3. Major scientific issues

This chapter specifies these major scientific issues related to oceanic plates and mantle, that can be addressed especially with the help of wide-band ocean-bottom / amphibious deployments and passive ocean floor seismology.

3.1 Lithospheric instabilities and upper mantle convection

![Fig. 4: Sketch of small scale sublithospheric mantle convection (SSC), so-called Richter rolls with their axes in the direction of plate motion. The onset of Richter rolls is possible only after the plate has cooled down and is growing thicker. SSC may explain the occurrence of intraplate volcanic chains and intraplate stress (modified after Ballmer, 2007)]](image)

Three distinct modes of mantle convection have been inferred from surface observations (e.g. Davis and Richards, 1992; Richter, 1973): plume mode (1), plate mode (2) and sublithospheric small-scale mantle convection (3, SSC). Plume mode convection is associated with hot spot volcanism (Morgan, 1981), plate mode with the creation and subduction of...
oceanic lithosphere, and SSC with instabilities of lithosphere older than about 80-90 Ma (e.g. Huang et al., 2003, see Fig. 4). For instance, SSC may be responsible for the undulation of the short wavelength (300 km) geoid in the Pacific (Buck and Parmentier, 1986). SSC was also suggested to explain the reduced subsidence of seafloor topography and constancy of heat flux at relatively old lithosphere (e.g. Parson and McKenzie, 1978), and later associated with the occurrence of non-hotspot volcanic chains (e.g. Pukapuka Ridges, Cook and Marshall island, Ballmer et al., 2007). The onset of SSC may be related to thermal lithospheric anomalies, and may be promoted by the age offset across fracture zones, and delayed by the existence of a thin asthenosphere (Huang et al., 2003). Other controlling factors for the onset and existence of SSC have been suggested, as for instance mantle plumes, large igneous provinces (LIP), or passive margins. High resolution surface wave studies have recently demonstrated that SSC in the oceanic mantle can be resolved in some cases (e.g. Ritzwoller et al., 2002, Montagner et al, 2002), but structural images are of limited resolution because seafloor deployments have not been available. Seafloor surface wave studies, which may be realized with the DEPAS pool, can improve the imaging of small scale upper mantle heterogeneities, and thus help to discover the scale and spatial distribution of SSC and therefore improve our understanding of the dynamic interaction of the oceanic lithosphere and underlying asthenospheric mantle. In this way it might also be possible to better understand the possible re-heating of the lithosphere (e.g. Zhong et al., 2007).

Several other studies focussed on anisotropy and flow patterns in the mantle beneath the oceans. Ekström and Dziewonski (1998) developed a global 3D tomographic model of shear-wave velocities in the Earth's upper mantle up to spherical harmonic degree 20. They concluded that the Pacific plate is the only large oceanic region, where radial anisotropy strongly deviates from the average anisotropy of PREM; vertically polarised shear waves in the mantle below the Pacific propagate faster than horizontally polarised ones. Further, the geographical variations in radial anisotropy beneath the Pacific plate are large and do not correlate with the age of the seafloor. They proposed that small asthenospheric convection in some areas of the Pacific plate, with length scales smaller than 1500 km, could cause the observed anomalies.

The plume mode convection, which is associated with hot spot volcanism, predicts significant upwelling of anomalously hot mantle when a plume head is arriving at the surface, leading to positive buoyancy, thick crust and sub-areal eruptions and the formation of large igneous plateaus. The largest igneous provinces in oceanic environment, the Ontong Java Plateau in the western Pacific, cannot be explained by such a mechanism, indicating that other convection modes than described by classical plate tectonic models are possibly more important that previously thought. The Ontong Java Plateau was formed at about 120 Ma in a submarine environment and not subaerially. It erupted anomalously low and now floats anomalously high (e.g. Korenaga, 2005). A working hypothesis explains the formation of the plateau by the dynamic entrainment of dense mantle rocks near a fast spreading mid-oceanic ridge, i.e. by small scale mantle convection. Sager (2006) and Neal et al. (1997) speculate that this, and other oceanic plateaus in the Pacific, seem to have formed near ridge-ridge-ridge triple junctions, and that such a tectonic situation promotes fast small scale convection and efficient mantle entrainment to form eruptive magmatic plateaus. Seismological studies of the mantle beneath the Ontong Java Plateau indicate indeed traces of anomalous upper mantle, but this has to be confirmed with other high-resolution regional studies. For instance, multiple ScS phases from a deep earthquake under the Solomon Islands reveal the existence of an anomalously slow root under the plateau (Gomer and Okal, 2003). Rayleigh wave studies
using island stations confirm the slow root and thick crust of the plateau, and additionally indicate a strongly developed low velocity zone at asthenospheric depths in the mantle (Richardson et al., 2000). Based on the splitting of direct and core-interacting shear waves recorded at temporary broadband stations on Micronesian islands, Klosko et al. (2001) suggest that the asthenospheric mantle flow beneath the Ontong-Java Plateau is diverted by the 300 km thick root of the Plateau. A similar pattern and behaviour is found beneath continental cratons. It is open whether other oceanic igneous plateaus have similar thick roots in the upper mantle and influence the asthenospheric flow, or whether the mechanism to form continental cratons and oceanic plateaus is more similar than previously thought. Ocean bottom broadband seismology may help to better unravel these questions.

3.2 Structure and state of the upper and lower boundary of the oceanic asthenosphere

Fig. 5: Location of the borehole seismic observatory in 5700 m depth, 560 m below seafloor (inverted blue triangle in A and B, B is zoomed version), together with piercing points for P-to-S (B, yellow crosses) and S-to-P (B, red crosses) converted waves at 70 km depth. C shows stacked radial P and S receiver functions, respectively. Only S-to-P piercing points sample the Parece-Vely basin because of the larger S-wave slowness. Converted signals from the Moho and the Lithosphere-Asthenosphere boundary show consistent structures with both P-to-S and S-to-P converted phases (from Kawakatsu et al., 2009).

Classical studies to reveal the depth, thickness and structure of the asthenosphere are based on S-wave models including S-wave anisotropy (see e.g. Gung et al., 2003 for a recent
inversion). These suggest that the Lehmann discontinuity, observed under continents and especially under old cratons, is a rheological boundary marking a transition from anisotropic to isotropic structure. The study of Gung et al. (2003) suggests that the varying depth of the anisotropic asthenospheric channel is marked by the Lehman discontinuity under continents at a depth of about 200-250 km, and the Gutenberg discontinuity under oceans at the depth of about 60 – 80 km.

Several body wave studies, mainly from island stations, indicate that the bottom side of the oceanic plate may be nearer to a sharp boundary than to a gradient zone (Li et al., 2004; Kumar et al., 2007). This is supported by global observations (Rychert and Shearer, 2009) and only recently by direct ocean bottom observations of S receiver functions (Kawakatsu et al., 2009, see Fig. 5). This implies, in contrast to the study of Gung et al. (2003), that the lithosphere-asthenosphere boundary is not a thermal boundary layer, but of chemical nature, e.g. due to rapid extraction of water at the onset of melting below the ridge (Hirth and Kohlstedt, 1996). It is however completely unclear, if this observation is of global relevance and the de-watering hypothesis needs to be tested against other possible mechanisms.

Temperatures in the mantle can be constrained by the mantle transition zone thickness (Anderson, 2007b). This is a first order parameter for global geodynamic models and is known only imprecisely especially beneath the deep oceans. Here, the only constraints are currently observations of precursors to free surface reflected waves, i.e. SS and PP, which have poor lateral resolution (Lawrence and Shearer, 2006). However, they clearly indicate a systematically thinner transition zone beneath the oceans compared to continents.

Fig. 6: Structural images of receiver function gathers resolving deep mantle structure as the slab (LBP) and the distortion of the 410 and 660 km discontinuity beneath Japan (from Tonegawa et al., 2006).

The potential of passive imaging methods to resolve mantle structure with unconventional data is shown by Tonegawa et al. (2006) where a novel receiver-function method was applied to image the upper mantle beneath the Japan subduction zone. Receiver functions are calculated from Hi-net tiltmeters by using stacked vertical components of broadband
seismograms as source time functions. The oceanic Moho and the lower slab boundary could be well resolved (Fig. 6). The slab could be traced down to a depth of 600 km. Uplift and depression of the 400 km and 660 km discontinuities are resolved in great detail, respectively. Teleseismic Mw > 5.5 earthquakes had been used from a period of 1 year recordings. They conclude that tiltmeter data are suitable to study the upper mantle by means of receiver functions. The tiltmeter recordings are very similar to the broadband seismograms in a frequency band between 0.02 and 0.16 Hz. This will be further discussed below.

Courtier and Revenaugh (2007) used ScS reverberations to search for mantle reflectors beneath the Tasman and Coral Seas. They identified a low velocity layer with its top at a depth of 352 km and associate this with partial melting in the deep upper mantle. Their method relies on high quality horizontal component recordings of local intermediate and deep focus earthquakes, and may be interesting when combing OBS with island station networks.

3.3 Stress in plates, basal drag and its relation to large inter- and intraplate earthquakes

![Fig. 7: Sketch to explain the difference of coupling between lithosphere and asthenosphere over a passive margin by basal drag (from Gung et al., 2003).](image)

The oceanic plates far from the subduction zones are occasionally sources of major earthquakes, which in many cases produced significant damage. An infamous example is the Lisbon earthquake of 1755. The great March 25, 1998 Mw 8.1 Antarctic plate earthquake is possibly the largest intraplate earthquake recorded instrumentally so far (Henry et al., 2000). Further famous examples are the 1929 Mw 7.2 Grand Banks earthquake in the North Atlantic Ocean (Bent, 1995) and the Ms 7.3 Baffin Bay earthquake (Bent, 2002). Additionally, the Indian Ocean has been particularly active with regard to intraplate events. The causes of these types of earthquakes and the formation of earthquake bearing regions in the interior or the boundaries of the oceanic plates is only partially understood (Wiens and Stein, 1985; Bird et al., 2002) and linked to the age of the oceanic lithosphere. The strength of the oceanic plate is strongly influenced by the heat production in the upper mantle. Steps in lithosphere thickness, typical along transform zones or fracture zones, may influence the coupling of the lithosphere to asthenospheric flow, which in turn may generate significant stress variations in the lithosphere.

Alternatively, lithospheric processes dominate in determining the stress in oceanic plates. For example, both the Indian Ocean and the region west of Iberia (source region of the Lisbon
earthquake) are characterised by diffuse deformation zones which accommodate minor relative motion between two tectonic plates (Africa and Europe west of Iberia, India and Australia).

Montagner (2007) and Gung (2003) studied the differences in radial and azimuthal anisotropies between oceans and continents, which might reflect a difference of coupling between lithosphere and asthenosphere through the basal drag (Figure 7). For instance, the lithosphere-asthenosphere coupling appears to be weak below most continental roots, in contrast to the Pacific plate at young ages, where the coupling is the first order effect in the uppermost 200 km for young ages. The study of anisotropy within the asthenospheric low velocity channel using DEPAS instruments has the potential to quantify better the relative contribution of basal drag versus gravitational stresses.

3.4 View in the mantle at triple junctions and the edges of plates

Another consequence of lateral variations in upper mantle temperature caused by unbalanced plate boundaries is the initiation of edge-driven convection in the upper mantle (e.g. King and Ritsema, 2000). This special type of small-scale convection has been postulated to occur mainly in the peripheries of old cratons bounded by oceanic plates (e.g. Africa, Bermuda, Greenland craton). The upwelling limbs of edge-driven convection may cause the thermal re-heating of the lithosphere and volcanism. Convincing seismological experiments to test the existence of edge-driven mantle flow beneath oceans have not been carried out so far, because they require ocean bottom deployments on a regional scale.

One tool to analyse the edge-driven upper mantle flow is again seismic anisotropy. For instance, the observed pattern of seismic anisotropy across the Hikurangi subduction zone, Central North Island, New Zealand, has been ascribed to the influence of trench parallel mantle flow beneath the slab and the presence of melt in the mantle wedge beneath the Central volcanic zone (Greve and Savage, 2009). Shear wave splitting observations near rapidly retreating slabs indicate the presence of plate driven flow in the asthenosphere near the South Sandwich Islands (Livermore, R., 2003).
4. Methods, novel approaches and logistical requirements

4.1 Rayleigh wave dispersion at the ocean floor measured from teleseismic earthquakes and ambient noise

Fig. 8: Example of Rayleigh wave dispersion (Z component, seismometer) observed on the Bear Island network of DEPAS stations (Sichuan Earthquake, 2008). Time (vertical axis) is in seconds and covers a range of 30 minutes.

Fig. 8 shows an example of dispersive Rayleigh wave trains observed with DEPAS instruments in the Barents Sea and on the North-Atlantic oceanic plate. The signal-noise ratio for long- and intermediate period Rayleigh waves is often good for ocean bottom stations. Therefore, passive deployments relying on surface wave studies have a high chance to be successful. This is also the experience made with other ocean bottom stations, and even broadband hydrophones may be used for classical one or two station dispersion studies (e.g. Tilmann and Dahm, 2008). However, deployment periods need to be longer compared to land stations.

On-shore, station networks are for a long time used to employ tomographic methods based on Rayleigh waves. The methods and approaches are continuously improving and further developed, for both, the classical earthquake analysis (a) and other more novel approaches as the impulse response derived from ambient noise (b).
A recent example for (a) is given by Levshin et al. (2007), where group velocity in the period range between 10 and 150 seconds was measured from broadband earthquake recordings from 1975 to 2005. A Monte Carlo inversion was used to construct a 3D shear wave velocity model of the crust and upper mantle beneath the Barents Sea down to a depth of 250 km. The thermal extension of the north-eastern Mid-Atlantic ridge system is identified in shear wave velocities (Fig. 9, left at the western border). The model could clarify the mantle expression of the continent-ocean transition in the Norwegian Sea and a deep high wave speed lithospheric root beneath the Eastern Barents Sea was resolved (Fig. 9, profile AA' and BB'). The lithospheric root is possibly a remnant of several Palaeozoic collisions, and the Barents Sea may be viewed as an old craton that is presently situated below sea level. The contemporaneous thickening of a high velocity body beneath the eastern Barents Sea basin is an intriguing indicator for the presence of mechanical coupling between crustal and upper-mantle structures, although the mechanism is not fully understood.

An example for the ambient noise method (b) is given by Shapiro et al. (2005), where a high resolution surface wave tomography was applied to 1 month ambient noise recorded at USArray stations. This study demonstrated that ambient noise tomography can improve the resolution and fidelity of crustal images obtained from traditional surface wave analyses. Other novel inversion methods, as the Eikonal tomography (Lin et al., 2009), argue to present a further improvement over traditional methods of ambient noise tomography.

The study by Yang et al (2008), where a two plane wave method was applied to teleseismic events to generate phase velocity dispersion maps for fundamental mode Rayleigh waves at periods from 25 to 100 s across the western U.S., verifies that at the short-period end of the spectrum (25-33 s) phase velocity maps are very similar to those derived from ambient noise. Both methods produce similar resolution at about the inter-station distance of 70 km at periods below 40 s.
The ocean floor is full of long and intermediate period ambient noise, which poses ideal conditions for ambient noise Rayleigh wave tomography. However, the local noise at periods > 25 s from stations at the ocean bottom poses a particular challenge (see study of Fan-Chi, 2006). The removal of tilt induced noise, by means of pressure signal correlated transfer functions, is possibly important for a successful ambient noise tomography studies on the ocean floor.

Another promising approach seems to combining the longer period (16–100 sec) fundamental mode Rayleigh-wave phase dispersion determined from teleseismic events with the short period (2-16 sec) fundamental and higher mode phase velocities from ambient noise (Harmon et al., 2007). The OBS experiment was conducted in the south central Pacific for a longterm deployment (6 month) of 10 broadband four-component OBS stations with an aperture of about 300 km and an interstation distance of about 100 km or more. Harmon et al. (2007) tried several processing techniques to improve the SNR for impulse responses from ambient noise, and finally conclude that the ambient noise analysis work reasonable for 2-16 sec period Rayleigh waves, but is difficult at longer periods. Second mode Rayleigh waves could also be included, which have peak sensitivity in the lower crust and upper mantle, and showed evidence for azimuthal anisotropy of about 2% of the S-wave velocity. Combining the ambient noise dispersion curves with phase velocities derived from teleseismic events, and
considering the aperture of the array, the authors could well resolve shear velocities to 45 km depth.

4.2 Medium aperture arrays (novel)

Dense seismological arrays enable the extraction of useful wavefield parameters like slowness and azimuth by stacking and frequency-wavenumber filtering, which are hard to obtain with single 3-component stations (Rost and Thomas, 2002; Schweitzer et al., 2002). Land based seismological arrays are since decades standard tools in seismology especially for the detection and identification of small events or tiny structure related phases. Examples are the identification of traveltime branches near the upper mantle triplication distance regions or steep incident core reflections (PcP) in the coda of direct P. Because stacking and frequency-wavenumber filtering methods can be applied, seismological arrays are very well suited to analyse the wavefield in noisy environments like the deep ocean (see Fig. 11).

Fig. 11: Slantstacks for the deep Mw 7.7 Kurile event of 5-Jul-2008 recorded at the broadband stations of the Bear Island network (courtesy F. Krüger). Direct and depth phases are marked, among them a very clear PcP (core reflection) observation.

Amphibian experiment with dense seismological arrays as an additional component may be beneficial if noise conditions are extremely difficult. This is for instance a severe problem if seismicity studies are envisaged. Often, only one or two OBS can be used for location if
noise is high. In this case, the complementary deployment of small-scale on-shore arrays may be helpful. These have high detection capabilities on array beams and give first estimates of a likely epicentre region, which, taken together with the OBS arrival time information can lead to robust earthquake location results, even under the mentioned difficult conditions.

The analysis of SKS fast splitting directions is a powerful tool to unravel mantle flow pattern beneath subduction zones or in regions where material is upwelling. Such information is for instance used to investigate whether crustal deformation is driven from below by slab retreat (e.g. Jovilet et al., 2009, for a study in the Mediterranean). Unfortunately, high quality S-, ScS and SKS phases are needed to resolve shear wave splitting, and convincing examples of splitting observed at ocean bottom stations are rare (see Fig. 12). Medium aperture 3-component arrays may help in this respect.

The logistical extra effort for the installation of temporary mid-aperture arrays is limited because typical interstation distances will be of the order of a few km up to a few tens of kilometres, what limits the extra needs for ship time. Small deviations of the actual station locations at the ocean bottom compared to the location of the water surface deployment positions bias results of standard array analysis only marginally and no additional logistical installation effort is necessary in that respect.

**Potential for ocean bottom receiver functions:**

Classical receiver function analysis is difficult for OBS data, because tiny signals on the horizontal components are used which are typically very noisy on OBS records. Therefore, either long recordings times (three years and longer) are needed, or, alternative techniques like a combination of landstations and OBS vertical recordings may be developed for S-
receiver functions. It has also to be mentioned that the orientation of the 3-component sensors on the ocean bottom is still a problem to make optimal use of the horizontal components. Tests to reach a better coupling are therefore necessary.

4.3 Combined analysis of different sensors (tilt, pressure, ground motion, EM)

It is well known that tilt-induced noise on horizontal components of OBS stations is often large at long periods. This tilting and wobbling of a free-fall station may be enhanced when the seafloor is unconsolidated, water-filled and soft ( mushy seafloor), and if seafloor currents are strong. If uncorrected, this poses a problem for techniques relying on horizontal components, as e.g. Receiver functions, SKS splitting or Love wave tomography. If the seismic sensor of a free-fall station is poorly leveled, the tilt-induced noise may even be transferred to the vertical component (e.g. Dahm et al., 2006). Another example of noise cross-coupling is observed for oceanic infra-gravity waves at long periods, which are measured on pressure sensors and induce compliance signals on vertical and horizontal seismometer components.

The combined analysis of the different sensors of an OBS station is a very promising approach to significantly improve the signal noise ratio, but is so far seldomly applied, unfortunately. Crawford and Webb (2000) and Stutzmann et al. (2001) have demonstrated that tilt noise on the vertical channel can be reduced by subtracting the cross-over signal predicted from the horizontal channels. An equivalent technique was used for removing the effect of infragravity waves (Webb and Crawford, 1999). Du et al. (2000) develop a data-adaptive polarization filter, using three component seismic channels, to reduce micro-seismic noise. The method is successful if the microseismic noise is stationary over periods longer than the periods of interest; an assumption that is often fulfilled at the seafloor. Other multicomponent techniques may be based on wavefield decomposition of passive ocean bottom data (e.g. Thorwart and Dahm, 2005) or on the combination of tilt measurements and seismic data.

Potential for Phase detection:
Both, the filter and the decomposition techniques are promising to improve the detection and analysis of body wave phases. Du et al. (2000) show several convincing examples for a large noise reduction in teleseismic recordings in Iceland, during periods of significant microseismic noise. Dahm et al. (2006) obtained an increase in signal noise ratio for teleseismic amplitudes about a factor of 200 for noisy OBS stations in the North Atlantic.

Potential for Receiver functions:
Receiver function studies are often masked by multiple reflections between waveguide layers, as for instance the crust. In the deep ocean, the water layer multiples pose a specific problem for short wavelength receiver function studies, since multiples arrive from above the station with strong amplitudes over 70 seconds or longer and make difficult the detection of tiny, converted phases from the upper mantle. The potential of a wavefield decomposition method, which is combining pressure and displacement sensors to separate the up- and downgoing wavefield, has been suggested promising for improving OBS receiver function studies (Thorwart and Dahm, 2005).
Another promising approach to improve the fidelity of horizontal channels and to help Receiver function and SKS studies on the ocean floor is to combine tilt and seismic data for noise reduction. Seismological broadband tilt data have rarely been recorded with OBS freefall stations. Onshore, however, Nishida et al. (2008) employed recently the potential of the Hi-net tiltmeters in Japan (679 stations) for ambient noise surface wave studies.

Dispersion curves of fundamental Rayleigh waves, fundamental Love waves and first overtones of Love waves were measured by means of fitting the synthetic cross spectra between pairs of stations within the circle. Frequency range between 0.05 to 0.2 Hz resolved sharp small-scale crustal and uppermost mantle anomalies to a depth of about 30 km. A novel receiver function method employing the Hi-net tiltmeters has already been discussed above (Tonegawa et al., 2006, see chapter 3.2)
Hensch (2009) describes a pilot OBS tiltmeter experiment conducted by the University of Hamburg in the Agean Sea (Hübscher et al., 2006) close to Santorini volcano. Four novel tiltmeter free-fall OBS stations were deployed in close proximity for a period of 9 month. Unexpectedly, the ocean bottom tilt traces showed a much better long period signal noise ratios than horizontal seismometer channels and were comparable to signals on island stations (Fig. 13.). We therefore see a large potential to improve the possibilities of receiver function and SKS splitting studies with DEPAS instruments, if tiltmeters are deployed together with seismic and pressure sensors in future ocean bottom experiments. The high resolution
tiltsensors are small, light and low-cost and may be mounted to the same gimbaling system as the broadband sensor of DEPAS stations, which would pose a minimal effort to realize such a novel measuring concept.

Potential of combined seismological – magnetotelluric measurements:

**Figure 14:** Left Panel: Station map of MT profile and derived electrical resistivity model. Middle Panel: Overlay of seismic velocities (Sallares et al. 2001) and location of interplate earthquakes denoted by blue circles and seismogenic earthquakes denoted by pink circles (from Dinc, 2008). Schematic in right panel: Arrow’s orientation and colours indicate pathway and different origins of fluids. The incoming plate consists of highly hydrated sediments on top of oceanic crust and mantle (pink layer on top of subducting slab), fairly hydrated oceanic crust and upper mantle. **Hydration:** Upon bending, deep faults are created allowing water to penetrate to mantle depths (C). Parts of water is stored freely in cracks and pore water; the other part is mineraly (chemically) bound in oceanic sediment, crust and (serpentinized) mantle. **Dehydration:** While subducting, water is released and expelled from cracks and pore water within the first few kilometres. Additionally, water released from early phase transition reactions in sediments is accumulated in shallow depths (D). Further in the subduction process, crust dehydrates. An effective dewatering of the hydrated mantle takes place in greater depths ~100 km due to De-serpentinization. An enigmatic feature (G) in the overriding plate before the volcanic arc hints to fluids.
Magnetotelluric measurements are an integral part of the geophysical data, since the mantle electrical conductivity derived from the data, is a well known complement to measurements of seismic velocity. Conductivity is sensitive to temperature, composition and hydration of the mantle, and therefore MT is widely used to provide constraints on mantle processes (Booker et al., 2004; Wannamaker et al, 2009, Kelbert et al., 2009). The potential of mantle scale MT measurements may be illustrated on the German large scale amphibious MT profile across the Costa Rican subduction zone, acquired in the framework of the SFB 574 „Volatiles and Fluids in subduction zones (Worzewski et al., 2009). The MT profile elucidates the subducting lithosphere, the more conductive overlying mantle wedge and conductivity anomalies associated with hydration and dehydration processes. Based on the MT profile, the entire hydration and dehydration cycle in the subduction zone could be imaged based on electrical resistivity variations. MT data has also been shown to be sensitive to flow induced anisotropy in a hydrated mantle (Evans et al., 2005), with the most conductive direction aligned with the a-axis of olivine alignment. This might be diagnostic, for example, of the radial flow pattern induced by the break-off of a instability at the base of the lithosphere. Similarities and differences between seismic and electrical anisotropy in concert with analysis of absolute values of conductivity and velocity have been used to identify compositional boundaries in the oceanic mantle. Although MT data provides important additional information on the oceanic mantle, the number of marine MT stations is limited. The marine EM group at IFM-GEOMAR/CAU Kiel lead by M. Jegen has now developed and built 30 broad band marine MT stations (10 through DFG funding and 20 through industry funding). The broad frequency range of the instruments make them especially adapt to mantle studies, where periods of up to 100 000 sec are essential. The need for these instruments is clearly demonstrated by the fact, that the instruments are in constant use in the last year and in the future two years in national and international research projects. The design of the instrumentation is such that it is compatible with the ocean bottom seismometer of the OBS pool operated at AWI.

5. Relation to other research programs and funding possibilities

- DFG Priority Programme SPP 1375, SAMPLE (www.sample-spp.de)

This SPP is focused on the study of the whole spectrum of phenomena related to continental breakup and passive margin formation and here mainly on crust and upper lithosphere of the passive continental margins in the South Atlantic region. The here planned initiative which is more focused on the open ocean and the lithosphere-asthenosphere system is therefore complement to this project.

- NEPTUNE Canada (neptunecanada.ca/research/research-projects)

NEPTUNE Canada is building the world’s largest cabled seafloor observatory off the west coast of Vancouver Island, British Columbia. The network, which extends across the Juan de Fuca plate, will gather data from a rich constellation of instruments deployed in a broad spectrum of undersea environments. The projects focus is the detailed analysis of a special important region (Cascadia subduction, Juan de Fuca plate system) with a wide variety of methods ranging from passive seismology (deep structure) to interactions within the ecosystem.
Dense regional array moving as leapfrog array across Northern America to determine how features seen at Earth's surface correlate with structural and compositional differences deep within the Earth. The focus is here mainly on deep continental structure and its imaging in unprecedented detail. Hundreds of broadband seismic stations are involved and mainly passive imaging techniques are used. Many new insights regarding plate tectonics beneath continents and at continental margins are to be expected as well as new imaging methods from which ocean bottom seismology in general will profit.

NERIES (www.neries-eu.org) is a Research Infrastructure project addressing observational seismology in a broad sense. Many elements within this Integrated Infrastructure Initiative European Commission-project are currently being realized, and a wide-scale collaboration with other projects within Europe and the US has been established. Broadband ocean bottom seismometer observatories were installed and are currently tested. NERIES and ESONET are organizing a workshop on the future of european ocean bottom seismology in February 2010 (Paris). The here proposed initiative will be presented there.

IODP/ICDP are international scientific deep drilling programs regarding deep ocean drilling and deep continental drilling. We will inform about our initiative on the ICDP/IODP workshop in March 2010 in Potsdam.

The US National Ocean Bottom Seismography Instrument Pool (OBSIP) organizes has been established by the National Science Foundation. Three facilities, located at Lamont-Doherty Earth Observatory at Columbia University, Institute of Geophysics and Planetary Physics at Scripps Institution of Oceanography and Woods Hole Oceanographic Institution, maintain and operate the instruments. First contacts have been established and a general mutual interest was expressed especially regarding common experiments in the Pacific region.

Ship time funding in Germany, proposal rules and deadlines:

1. **Mittelgrosse Forschungsschiffe** (z.B. Nordatlantik, Mittelmeer, Nord- und Ostsee, siehe www.io-warnemuende.de/mittelgrosse-forschungsschiffe.html): **Mitte Januar**


6. Proposals suggested by September 2009 (open list)

A. Studies with focus on specific regions

A.1: South Sandwich subduction zone (marginally studied intra oceanic subduction system characterized by rapid slab retreat, contact of oceanic asthenosphere with Antarctic continental keel, strong mantle upwelling in the East due to Bouvet hotspot; see 3.1, 3.2 and 3.4). Suggestion from AWI, Uni Frankfurt, Uni Potsdam, GFZ Potsdam, Cambridge (local earthquake tomography, surface wave tomography, shear wave splitting analysis, receiver functions). 36 OBS in combination with island based seismological arrays and single stations.

A.2: Seismicity analysis of the Southwest-Indian Ridge (SWIR) for better understanding of active tectonic and magmatic processes at ultraslow-spreading ridges. Suggestion from AWI, see also 3.3 and 3.4, 16 OBS, 3 months in 2012.

A.3: Interdisciplinary research project to study the deep structure and genesis of the Ontong Java plateau, Suggestion from Univ. Bochum; see also 3.2.

A.4: Deep structure and microseismicity of the North and Baltic Sea (AWI).

A.5: Structure of the Northern Anatolian Fault Zone with focus on the Maramara Sea region (amphibian experiment (Univ. Kiel); see also partially 3.3.

A.6: Dynamic topography of the Eltanin and Udintzев transform systems to study coupling between topography/tectonics and deep ocean currents, crustal structure and strain rates from local seismicity (AWI); see also 3.3.

A.7: Origin and causes of intraplate seismicity in the corridor between Barbados subduction zone and Cape Verdes, Central North Atlantic. A consequence of relative motion between North and South America? (IFM Geomar)

B. Studies more focused on special themes.

B.1: Fine structure of the oceanic upper mantle from surface wave analysis studied with a wide OBS network, here suggested for the Pacific, South Atlantic region (Uni Bochum); see 3.1, 3.2, 3.4.

B.2: Better coverage of the southern hemisphere to fill gaps in global tomography studies by installation of quasi-permanent mini-arrays along regularly visited ship routes for instance in direction of the Neumayer III base or realized as moving larger arrays (AWI).

B.3: Study of the influence of the Island plume on the mantle structure in the Northern Atlantic. Mapping of the crust and mantle structure of the whole European oceanic plate with moving large aperture arrays consisting of 12-24 stations or one very large array of up to 80 stations. Perhaps realized as EU initiative (AWI); see also 3.1, 3.2, 3.3 and 3.4.
B.4: Time development of hotspots and their tracks in the oceanic lithosphere using the example of the Rana Fracture Zone, Tristan da Cunha, Mt. Cameroon, AWI, see 3.2 and 3.4.

B.5: Seismotectonics and mantle shearing at major conservative plate boundaries, like Azores-Gibraltar fault zone, Panama Fracture zone (IFM-Geomar)

B.6: Mantle wedge hydration at subduction zones (IFM-Geomar)

It should be mentioned that all groups signalled that they are open for piggy-back experiments to contribute data for global studies.

7. References


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