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Zühlsdorff, L.

HIGH RESOLUTION MULTI-FREQUENCY SEISMIC SURVEYS AT THE EASTERN JUAN DE FUCA RIDGE FLANK AND THE CASCADIA MARGIN -EVIDENCE FOR THERMALLY AND TECTONICALLY DRIVEN FLUID UPFLOW IN MARINE SEDIMENTS



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Reports can be ordered from:

Gisela Boelen

Sonderforschungsbereich 261

Universität Bremen

Postfach 330 440

D 28334 BREMEN

Phone: (49) 421 218-4124

Fax: (49) 421 218-3116

e-mail: eggerich@uni-bremen.de

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Zühlsdorff, L.

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Chapter 1: Introduction

The hydrologic character of the earth's crust plays an important role in a number of geological and geophysical processes, which occur in a variety of geological settings as oceanic spreading centers and convergent margins. At spreading centers and flanks of submarine volcanic ridges, large-scale hydrothermal circulation significantly affects the heat flow as well as the global geochemical cycles of both lithosphere and oceans [e.g. *Lowell et al.*, 1995]. At subduction zone accretionary prisms, the presence of water strongly influences the deformation of sediments and the development of pore pressures in excess of hydrostatic [e.g. *Moore and Vrolijk*, 1992; *Carson and Screaton*, 1998]. Furthermore, fluid flow and sediment dewatering are associated with the formation of mineral deposits and the migration and entrapment of hydrocarbons. One example is the rise of methane and the formation and subsequent dissociation of methane hydrates, which are important as a possible energy source and may affect atmospheric composition and global climate change [*Kvenvolden*, 1988; *Kvenvolden*, 1994].

However, a quantitative description of fluid flow has often been difficult. This is partly due the fact that porosity, permeability, and consequently pore pressure are time variable rather than static [e.g. Waldner and Nur, 1984], and fluid expulsion may be coupled to episodic fault displacement and the earthquake cycle [Moore and Vrolijk, 1992]. Another reason is that information about the relative importance of focused flow versus diffuse flow is rarely available [e.g. Suess et al., 1998]. Upflow velocities along faults were suggested to be two to three orders of magnitude higher than dispersed flow through the sediment [Moore et al., 1990], but dispersed flow, which is accommodated by intergranular and small scale fracture permeability, leaks out over much larger areas and thereby causes an unknown proportion of water loss [Carson and Screaton, 1998].

Valuable information about upward fluid migration has been provided by observations of the physical and chemical consequences of fluid flow as the alteration of both rocks and fluid or the creation of local thermal or chemical anomalies. Common methods to investigate submarine fluid discharge include heat flow measurements [e.g. *Davis et al.*, 1990], chemical pore fluid analyses on sediment cores [e.g. *Wheat and Mottl*, 1994], and sampling and mapping of early diagenetic deposits near the seafloor [e.g. *Kulm and Suess*, 1990]. Furthermore, packer experiments represent a first step to record the

modulation of fluid flow over time at the location of a drill site. The circulation obviation retrofit kit (CORK) minimizes the thermal and chemical effects on the formation from drilling-induced disturbances, samples fluids, and monitors in situ thermal and hydrological conditions long after drilling disturbances have dissipated [*Davis et al.*, 1992b].

Seismic data can contribute to the investigation of fluid upflow if flow-related processes affect sediment structure or physical properties. Since both seismic velocity and bulk density of marine sediments strongly depend on porosity, porosity anomalies are often associated with velocity anomalies or variations in seismic reflection amplitude, indicating compaction driven fluid expulsion [*Bray and Karig*, 1985; *Minshull and White*, 1989], elevated fluid pressures [*Waldner and Nur*, 1984; *Calvert and Clowes*, 1991], or the presence of permeable pathways (Chapter 4). Seismic data have also been used to identify other potential fluid conduits like faults and to map the occurrence and distribution of natural gas hydrates and seafloor cemented layers (Chapter 6), which may be diagnostic of upward fluid flow [e.g. *Hyndman and Davis*, 1992; *Carson et al.*, 1994]. However, to image flow-related features and to estimate the spatial extend of potential upflow zones and the volumes of sediment and pore space that host the flow, the quality and resolution of seismic data, which both depend on the source frequency range, have to be sufficiently high.

One of the characteristics of seismic data is that signal penetration decreases if higher source frequencies are used to image sediment structures in greater detail. Furthermore, seismic data are often deteriorated by other frequency dependent effects like scattering or interference. These difficulties can be overcome if a multi-frequency data set is acquired, which provides maximum temporal resolution at different depth levels and is sufficient to investigate a possible frequency dependence of seismic features.

To meet these requirements, all components of the multi-channel seismic system of the University of Bremen, including seismic sources, streamer, and the data acquisition unit, were designed to optimize lateral and vertical resolution. Furthermore, the system allows to operate different seismic sources in an alternating mode. During R/V Sonne Cruise SO111, which was carried out at the eastern Juan de Fuca Ridge flank and the Cascadia margin in late summer 1996 [Villinger et al., 1996], the system was first employed and tested using a GI-Gun (100-500 Hz) and a water gun (200-1600 Hz) by turns along each seismic line. The seismic records were further combined with hydroacoustic data as provided by the narrow beam Parasound echosounder (4000 Hz) and the Hydrosweep multibeam swath sounder, which mapped the seafloor bathymetry. Since seismic and hydroacoustic systems could be operated simultaneously due to the different source frequency ranges, required ship time was minimized and the accuracy of navigation was improved. Thus, a unique, high resolution multi-frequency data set was provided, which could in some areas be supplemented by previously collected single channel data of lower frequency content and resolution.

The working area of Cruise SO111 is appropriate to study different types of fluid upflow in greater detail. Massive turbidite input from the continent into the Cascadia Basin (Figure 1.1) provides a low permeable sediment cover above young and permeable crust, which is extremely smooth in some areas and rough in others. Sediments lap onto the eastern Juan de Fuca Ridge flank and create a sharp boundary between sediment-free and sediment-covered crust just 18 km east of the spreading axis [e.g. Davis et al., 1992a]. Near this boundary as well as at some locations further to the east, Ocean Drilling Program (ODP) Leg 168 drill holes penetrate the sediments and upper basement. The project was targeted to elucidate the fundamental physics and fluid chemistry of ridge-flank hydrothermal circulation and the consequent alteration of the upper crust and sediments [Shipboard Scientific Party, 1997a]. During Cruise SO111, a high resolution seismic survey was carried out in the vicinity of the western ODP Leg 168 sites (Figure 1.1), shortly after the leg was finished. The objective was to identify zones of thermally driven fluid upflow through the sediments as were previously predicted by Wheat and Mottl [1994], to investigate the shape and extent of these zones, and to image the associated sedimentary structures.

Approximately 3.5 km of Cascadia Basin sediments are later scraped off the downgoing Juan de Fuca plate, which is subducted along the Cascadia margin [*Davis and Hyndman*, 1989; *Hyndman and Davis*, 1992]. As a consequence, a large sediment wedge is formed, which is associated with a rapid decrease in porosity with depth and the tectonically driven expulsion of large volumes of fluid [e.g. *Davis et al.*, 1990]. ODP Leg 146 was directed to the investigation of this fluid flow and its relation to the deformation of sediments within the accretionary prism [*Shipboard Scientific Party*, 1994]. During Cruise SO111, a limited high resolution seismic survey was carried out in





the vicinity of the northern ODP Leg 146 sites (Figure 1.1) to gain insight into the formation of gas hydrates and the regional distribution of diffuse and confined fluid release.

The objectives of this study are to present the results of both seismic surveys, to demonstrate how high resolution seismic data can contribute to the investigation of fluid flow, and to discuss the additional information provided by a multi-frequency data set. The text is organized as follows:

First, seismic data acquisition and processing are described with an emphasis on the aspects of high resolution (Chapter 2), followed by a brief introduction to fluid upflow through marine sediments (Chapter 3).

In Chapter 4 (published in *J. Geophys. Res.*, 104, July 1999), a frequency-dependent decrease in seismic reflection amplitude is presented, which occurs above a buried basement ridge on the eastern Juan de Fuca Ridge flank. Since the amplitude anomalies are found to be associated with zones of higher porosity and permeability, and independent evidence suggests that these zones are correlated with thermally driven fluid upflow, it is inferred that fluid discharge above the basement ridge can be imaged and mapped by high resolution seismic data.

To ground truth the seismic records, traces were correlated with synthetic seismograms calculated from ODP Leg 168 core logging data (Chapter 5, to be submitted to *J. Geophys. Res.*). It is shown that seismic reflections are related to variations in GRAPE density, which is also sufficient to model the frequency-dependent decrease of seismic reflection amplitudes above the buried basement ridge. Furthermore, a combined interpretation of seismic and borehole data provides geologic information as relative accumulation rates and a distinction between different depositional regimes in the vicinity of the basement ridge.

In Chapter 6 (accepted for publication in *Geologische Rundschau*, April 1999), information about fluid discharge at the northern Cascadia accretionary prism is provided by mapping the occurrence and distribution of both, thin zones of high near seafloor reflectivity, which may be associated with the precipitation of methane derived carbonates, and a methane hydrate bottom simulating reflector (BSR). It is shown that at least two different types of episodic fault-controlled fluid release superimpose the known diffuse fluid seepage at the northern Cascadia margin.

In the final Chapter 7, the results of this study are summarized. It is further proposed how the new insights can potentially be used to constrain future investigations and models in order to quantify flow-related processes.

Chapter 2: High resolution multi-channel seismic data

2.1 Introduction

In order to collect seismic reflection data in a marine environment, pressure waves are generated, which travel down through the water into the earth. A proportion of the source energy is reflected where changes of seismic impedance, i.e. the product of bulk density and seismic velocity, are present. The returning echoes are received by hydrophones and then digitally recorded on tape.

Source signals are usually produced by the rapid acceleration of water by a water gun, which generates an acoustic signal of implosive type, or by the impulsive release of highly compressed air from an air gun. Guns with chambers of smaller volume provide higher maximum frequencies and broader frequency ranges, and thus source signals of shorter length. However, when source frequencies are high, also the receiver array and the data acquisition unit should be designed to optimize temporal and lateral resolution. In addition, a very careful seismic processing is required to gain maximum data quality and to preserve the structural information.

In this chapter, both acquisition and processing of the high resolution multi-channel data collected during R/V *Sonne* Cruise SO111 [*Villinger et al.*, 1996] are described and data properties are discussed.

2.2 Data acquisition

The basic components of the GeoB marine multi-channel seismic system include different seismic sources, which can be operated in an alternating mode, the streamer carrying the hydrophones, the streamer control unit, and the data acquisition unit controlling data management and recording (Figure 2.1). All components were designed to meet the requirements of high resolution.

2.2.1. Seismic sources

During Cruise SO111, a Sodera S15 water gun with a chamber volume of 0.16 L and a GI-Gun (Generator/Injector-Gun) with reduced volumes of both chambers of 0.4 L were used by turns. For both guns, the far-field source signatures and corresponding spectra under optimum conditions are shown in Figure 2.2.

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Figure 2.1 Acquisition of multi-channel seismic data. Hull-mounted hydroacoustic systems as Parasound and Hydrosweep can be operated simultaneously.





Figure 2.2 Water gun and GI-Gun far-field source signatures and spectra (unfiltered) under optimum conditions as measured by the manufacturer.

A typical water gun signal is characterized by a precursor, i.e. an early arriving portion of the source signal, which is best observed shortly before seafloor reflection time (compare to Figures 2.2 and 2.7a). However, the amplitude of the precursor is much smaller than the amplitude of the main signal and does not significantly hamper the quality of the seismic records.

In contrast, a conventional air gun produces a bubble pulse with a high amplitude, which follows the source signal and is difficult to remove from seismic records by processing. Instead, a GI-Gun could be used, which generates the source signal by releasing the generator chamber first and then significantly reduces the bubble pulse by a time delayed release of the injector chamber.

The alternating release of the sources and the time delay between generator and injector are controlled by a PC based trigger unit.

2.2.2 Streamer

The hydrophones are carried by an oil-filled streamer of 300 m length (Syntron Inc.) and are organized into groups and subgroups. The signal of a receiver group or subgroup is stacked to enhance the S/N (signal to noise) ratio at each channel during data acquisition. However, long hydrophone groups suppress high source frequencies at greater inclination angles since the differences in travel time between the near shot and far shot receivers of the group may cause destructive interference. Especially if the water depth is small, very short groups or even single hydrophones may be required to provide sufficient receiver response and bandwidth. Thus, the streamer was designed to allow for separately programmable hydrophone subgroups, which can be optimized for a given target and the required resolution. In the Cascadia Basin, where water depth is > 2000 m, 24 groups with a length of 6.25 m and at a spacing of 12.5 m were used.

Depth control of the receivers was provided by four cable levelers (birds), which kept the streamer depth within a range of 1 m. In addition, magnetic compass readings allow the determination of the position of each streamer group relative to the ship's course. The birds are remotely controlled by a PC based bird control unit (MultiTRAK, Syntron Inc.) onboard the ship.

2.2.3 Data acquisition unit

Data management and recording was controlled by the data acquisition unit (Bison Spectra), which was specially modified to work at high shot and sampling rates. During

the surveys in the Cascadia Basin, data were sampled at 125 µs and shots were recorded every 10 s. Accordingly, in the alternating mode, the time interval between shots of the same source was 20 s. Data were stored demultiplexed in standard SEG-Y format, which consists of a reel header and the data records, each of them including 240 bytes of trace header information as sampling rate, number of trace samples, shot number, channel number, and recording time. Almost every seismic processing software package, which is available on different computer platforms, is capable of reading this format.

2.3 Processing of multi-channel seismic data

Standard seismic processing as filtering, stacking, and migration was carried out with the Seismic Unix (SU) software package [*Stockwell*, 1997]. A fundamental part of the processing flow is the common depth point (CDP) stack, which increases the S/N ratio, improves the data quality, and allows an estimate of subsurface seismic velocity. However, CDP processing requires trace header information about source and receiver location, CDP number, and static correction time. With regard to the high frequency content of the data, this information can only be provided by additional software, which takes possible temporal and spatial variations of both streamer depth and streamer heading into account. Thus, FORTRAN codes were developed to utilize all available bird information in combination with the ship's navigation data.

2.3.1 Procedure

To create a list of the required trace header values, four files of input information have to be provided: (1) A list of shot number and recording time as generated by the data acquisition unit, (2) a navigation file with the ship's course, heading, and geographic position at a sample interval of at least 20 s, (3) depth records of all cable levelers, which were sampled every 10 s by the bird control unit, and (4) a similar file of the bird control unit providing magnetic compass readings along the streamer. To facilitate a combination of the different files at a specific time, a linear interpolation was applied to resample all information to a time interval of 1 s. Then, the ship's geographic position as well as the depth and the magnetic heading of the cable levelers can be determined for each shot time. For simplification, the ship's position was defined to be identical with the position of the seismic sources. This causes a constant offset since the receiver of the DGPS (Differential Geographic Positioning System) was located at the ship's bow whereas the sources were towed behind the ship. However, the relative distances between sources and receivers are not affected and a constant lateral shift in CDP positions can later be corrected, e.g., when borehole data are compared to seismic records.

To determine the receiver positions, the compass readings of the cable levelers are interpolated along the streamer to provide magnetic heading values at an interval of 1 m. Further, a local magnetic declination is estimated to convert magnetic heading into geographic heading. Thus, starting at the source position and taking the ship's course as a reference, possible curvatures of the streamer are successively reconstructed from variations of heading values. The main advantage of this procedure is that the receiver positions can even be determined for each channel when the streamer or parts of it are not straight, e.g., when the ship is turning or the streamer is drifting sideward due to currents.

The main task of the software is the definition of CDP numbers. As a first step, the midpoint between the source and receiver location is determined for each trace. If the streamer is straight and its heading is consistent with the ship's course, all midpoints should be distributed along a straight line. Furthermore, if source and receiver locations coincide in a regular manner, the midpoints are expected to cluster at CDP positions. In reality, however, the ship's course and speed may vary, the streamer may be drifting, and midpoints are scattered around the desired seismic line (Figure 2.3). To account for this, a CDP is defined to cover a small area (CDP bin) with the meaning of a logical container in which a number of midpoints may be grouped. The circular bins have a defined size and are evenly spaced along the seismic line, which has not necessarily to be straight. When the bin size is chosen to be large and the distance between bins is sufficiently small, bins may even overlap. However, each midpoint is assigned only once to a bin, based on the minimum distance to nearby bin centers. Midpoints which are not located within any bin are marked with a zero CDP number, and the corresponding traces will then be ignored for further processing (Figure 2.3).

As a result, the number of a CDP bin, the source position, and the receiver position are determined for each shot and recording channel. The next step is the definition of a



Figure 2.3 Midpoints (dots) between source and receiver locations for a number of successive shots. The CDP bins (circles) at a spacing of 20 m are slightly overlapping. When the ship is turning, the streamer is not straight and the midpoints are scattered around the seismic line. Consequently, the number of midpoints, which can be grouped within each bin, decreases.

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static correction time for each trace to account for lateral and temporal depth variations of seismic receivers. For this purpose, the depth readings of the cable levelers are interpolated along the streamer to provide a depth value at each receiver position. Thus, a static correction time for each trace can be calculated from a reference depth and the seismic velocity of seawater.

Finally, the static correction time as well as CDP number and geographic positions are assigned to the trace headers of each seismic record. The main data processing sequence including filtering, static correction, normal move-out (NMO) correction within CDP gathers, CDP stack, and migration can then be carried out using the standard modules of the SU software.

2.3.2 Accuracy of navigation processing

Before receiver positions can be calculated from the compass readings of the birds, magnetic heading has to be converted into geographic heading, which requires knowledge about the local magnetic declination. If magnetic declination estimates are incorrect, the error in calculated receiver positions will grow with increasing offset. This effect can not be distinguished from a real sideward drift of the streamer, which may be caused by surface currents. Therefore, a rough estimate of magnetic declination as provided from maps appears to be insufficient. Instead, software from the National Geophysical Data Center in Boulder, Colorado, was used, which estimates magnetic declination from the world magnetic model and takes spatial and temporal variations into account. In some rare cases, however, deviations of up to 4° from the correct value may still occur in the presence of extreme local magnetic anomalies. This would result in an error in geographic position of about 28 m for the far offset receiver of a 400 m long streamer. In the Cascadia Basin, however, the observed streamer heading was usually consistent with the ship's course and the accuracy of magnetic declination estimates was assumed to be sufficient.

The quality of the CDP stack strongly depends on the accuracy of static correction and NMO-correction. Especially for high temporal resolution, even slight variations in reflection time may result in destructive interference of the signal and a subsequent significant loss of information and energy. Furthermore, the SEG-Y format, which is still used by standard seismic software, is not designed for appropriate treatment of high resolution data. As an example, the minimum static correction time, which can be

assigned to a SEG-Y header, is 1 ms. Although this is sufficient for most source frequency ranges of conventional seismic data, smaller corrections may be required for source frequencies above 1000 Hz. As a solution, the calculated static correction time as well as the sampling rate was multiplied by a factor of 100, allowing to take two more digits into account. The static correction was then applied, and the sampling rate was subsequently set back to the original value.

For the data collected in the Cascadia Basin, static corrections were almost negligible since temporal variations of the depth readings appeared to be very small and the standard deviation from a target streamer depth was calculated to be less than 0.5 m at the location of the cable levelers. However, a straight forward static correction as described above may not be sufficient if both streamer motion and depth readings are affected by irregular wave movement. In this case, depth readings will scatter around the desired values and the subsequent NMO correction may result in destructive interference.

For NMO correction, a seismic velocity model is needed. In the Cascadia Basin, the streamer length was short compared to the water depth and an error of about 10% was expected for estimated velocities, even if traces from several adjacent CDP gathers are combined. However, observed velocity variations appeared to be small and a constant velocity could be assumed for all seismic lines. This is in good agreement with velocity measurements on cores from various ODP sites in the study area.

2.4 Properties of multi-channel seismic data

2.4.1 Trace spacing and S/N ratio

The S/N ratio of noisy seismic data increases with the number of stacked traces, and thus, with the length of hydrophone groups (compare to section 2.2.2) as well as with the size of CDP bins. The maximum bin size is defined by the size of the first Fresnel zone, which depends on source frequency and water depth. If the bin size exceeds this limit, destructive interference of stacked seismic amplitudes may cause a significant reduction of data quality. However, if bins are chosen to be smaller than the first Fresnel zone and the distance between the bins, i.e. the lateral trace spacing, is reduced, CDP processing consequently produces a higher number of traces along the seismic line. This may be favorable since migration of high frequency data is often hampered by aliasing

when the trace spacing is too large and the seafloor is inclined. Furthermore, a visual lateral correlation of seismic reflections is facilitated at smaller trace spacing. However, if the number of bins increases, the number of traces within each CDP gather decreases and the S/N ratio may not be sufficient. Therefore, bin size and bin spacing have to be chosen carefully, especially when the number of available channels is limited. Two examples of what may be a good compromise between trace spacing and S/N ratio are shown in Figure 2.4 for the Cascadia Margin. In the Cascadia Basin, where topography is generally moderate, a bin spacing of 20 m was chosen for most seismic lines, which corresponds to a fold of nine traces for each bin.

2.4.2 Frequency content

Typical amplitude spectra of recorded traces show that the source frequency ranges of GI-Gun and water gun are approximately 100-500 Hz and 200-1600 Hz, respectively (Figure 2.5). For a streamer depth of 5 m and a seawater velocity of 1500 m/s, notches at harmonics of 150 Hz may be explained by ghost reflections from the surface, which have negative polarity and follow so closely behind the primary pulse as to interfere destructively.

Application of different band-pass filters to GI-Gun and water gun data indicates, that the complete frequency bandwidth is required to produce the high resolution seismograms presented in Figures 2.6a and 2.7a. If the higher frequencies are neglected, temporal resolution decreases significantly (e.g. in Figure 2.7b at 3.6 s) and the traces are affected by ringing, even if the filter flanks are moderately tapered (Figure 2.6b). In contrast, if the filter cuts off the lower frequency range, the remaining bandwidth appears too small to produce single sharp reflections (Figures 2.6c and 2.7c).

2.5 Combination of seismic and hydroacoustic data

To provide a unique multi-frequency data set, multi-channel seismic data can be combined with both, hydroacoustic data of even higher resolution and single channel seismic data of lower frequency content.

2.5.1 Hydroacoustic data

The Parasound echosounder (STN Atlas Elektronik) provides detailed acoustic images of the upper 50-100 m of the sediments. Due to the high signal frequencies of 2.5-5.5 kHz and the narrow beam angle of 4°, diffraction hyperbolae are mainly suppressed and



Figure 2.4 Migration of Line Geob96-069 (GI-Gun) at the Cascadia Margin. (a) A trace spacing of 10 m corresponds to a fold of 3. (b) A trace spacing of 20 m corresponds to a fold of 9. At the smaller trace spacing, inclined structures (e.g., at CDPs 350-480) are better imaged, whereas the S/N (signal to noise) ratio is higher at the higher fold. Vertical exaggeration (VE) was calculated for a constant velocity of 1700 m/s.



Figure 2.5 A stacked seismic trace of (a) GI-Gun and (b) water gun data in the frequency and time domains. Low frequency noise was eliminated by high-pass filters. Source frequency ranges are 100-500 Hz and 200-1600 Hz, respectively.

very high temporal and lateral resolution can be achieved. Furthermore, heave, roll, and pitch are fully compensated. The ParaDigMa system [*Spieß*, 1993] digitizes the data and stores them in SEG-Y format for further processing. In addition, the Hydrosweep multibeam swath sounder (STN Atlas Elektronik) maps the seafloor bathymetry and shows small-scale features at the seafloor. Both hull-mounted hydroacoustic systems are available on the German research vessels Sonne, Meteor, and Polarstern.

Due to the different source frequency ranges, it is possible to record seismic and hydroacoustic data simultaneously (Figure 2.1). This improves the accuracy of navigation and minimizes the required ship time.

2.5.2 Single channel seismic data

In some areas, supplementary single channel seismic data were available, which were collected onboard RV *Tully* in 1995 (compare to Chapter 5) and during RV *Sonne* Cruise SO111 in 1996 (compare to Chapter 6). In both cases, all 16 hydrophone groups of a 100 m long streamer (Teledyne) were stacked and a shot rate of 10 s corresponded to a shot spacing of 25 m. However, different GI-Guns were used, which provided source frequency ranges of 20-120 Hz (RV *Tully*) and 50-150 Hz (RV *Sonne*), respectively. Single channel and multi-channel seismic data were partly collected along the same lines across the drill sites of ODP Leg 168.

2.5.3 The multi-frequency data set

The custom software described above provides exact geographic CDP positions, which can be compared to the ship's navigation data, and thus, to the recording time of the Parasound system. In addition, shot positions of the single channel data are available from the trace headers. Therefore, single channel seismic, multi-channel seismic, and hydroacoustic data can be combined despite of differences in data acquisition, shot rate, and trace spacing.

As shown in Figures 5.9a and 5.9b, single channel data (20-120 Hz) and multi-channel GI-Gun data from the Cascadia Basin are sufficient to reveal the basement morphology beneath a 150-200 m thick cover of turbiditic sediments. However, as observed for water gun and Parasound records, signal penetration decreases with increasing resolution, but sediment structures are imaged in greater detail (Figures 5.9c and 5.9d). To utilize the complete seismic information available, all data are combined for an



Figure 2.6 CDP-stack of Line GeoB96-078 (GI-Gun) after application of different band-pass filters: (a) 100-2000 Hz, (b) 100-240 Hz, and (c) 240-2000 Hz (taper not included). For comparison, an amplitude spectrum is given in Figure 2.5a. Vertical exaggeration was calculated for a constant velocity of 1500 m/s.



Figure 2.7 CDP-stack of Line GeoB96-078 (water gun) after application of different band-pass filters: (a) 200-2000 Hz, (b) 200-800 Hz, and (c) 800-2000 Hz (taper not included). For comparison, an amplitude spectrum is given in Figure 2.5b. Vertical exaggeration was calculated for a constant velocity of 1500 m/s.

integrated interpretation. Furthermore, the multi-frequency data set provides the opportunity to investigate a possible frequency dependence of seismic features.

Chapter 3: Fluid upflow in marine sediments

3.1 Introduction

The flow of fluids through marine sediments is related to a variety of geologic processes, which have been of considerable scientific, economic, ecological, and societal interest in recent decades. Among many others, these processes include the maturation, migration, and entrapment of hydrocarbons, the formation of gas hydrates and mineral deposits, and the physical and chemical alteration of crust and sediments due to ridge flank hydrothermal circulation.

Since fluid migration is such a broad subject, this chapter is not intended to present a complete description of the various types of fluid flow and the way in which they are significant. It is rather confined to some aspects related to the hydrothermally driven fluid expulsion at the eastern Juan de Fuca Ridge flank and the compaction driven dewatering at the northern Cascadia accretionary prism. More detailed information about the specific situation in both study areas is provided at the beginning of Chapters 4 and 6, respectively.

3.2 Fluid supply

Processes as mineral precipitation, diagenesis, or the transport of heat and chemicals, are related to the flow of various types of fluids. These include liquid or gaseous thermogenic fluids, biogenic gas, volcanic or hydrothermal gas, or pore water [e.g. *Hovland and Judd*, 1988]. Ridge flank hydrothermal circulation is believed to involve basically seawater which alters the composition of the crust and is, in turn, altered by itself by a two-way chemical exchange process [e.g. *Lister*, 1972; *Hartline and Lister*, 1981; *Lowell et al.*, 1995]. The dewatering of accretionary prisms commonly involves mainly pore water as well as dissolved or free gaseous hydrocarbons [e.g. *Hyndman and Davis*, 1992; *Moore and Vrolijk*, 1992; *Carson and Screaton*, 1998].

The mobility of gas depends on parameters like solubility, density, and molecular size. Of the smaller gaseous hydrocarbons methane has the smallest size and is most mobile [e.g. *Hovland and Judd*, 1988]. In shallow marine sediments commonly two types of methane may be found. The first is of thermogenic nature and has migrated upwards from mature source rocks and reservoirs. The second is of biogenic nature and is produced at depths shallower than about 1000 m. Both types can be distinguished by their isotopic composition [*Claypool and Kaplan*, 1974].

If sufficiently large volumes of gas are supplied, natural gas hydrates (clathrates) may form at high pressure and low temperature [e.g. *Sloan*, 1990; *Paull et al.*, 1994; *Kvenvolden*, 1994]. Gas hydrates are crystalline ice-like compounds composed of water and gas which can store much more gas than an open gas-filled sediment pore system [e.g. *Kvenvolden*, 1988; *Hovland and Judd*, 1988]. The most common hydrate forming gas is methane [e.g. *Claypool and Kaplan*, 1974; *Dillon et al.*, 1994], and thus gas hydrates may provide an enormous methane reservoir in marine sediments. If temperatures increase or the sediments become depressurized, the decomposition of gas hydrates may be considered as an efficient secondary source of methane [e.g. *Dillon and Paull*, 1983; *Kastner et al.*, 1995]. The released gas can subsequently migrate upward to form new hydrate, to be trapped beneath low permeable barriers, or to be expelled at the seafloor.

3.3 Vertical fluid transport at hydrostatic pore pressures

The different transport mechanisms of gaseous hydrocarbons through geologic units include diffusion of gas molecules, solution in migrating pore water, and direct bubble movement through permeable layers and vertical zones of weakness. Some authors state that gas is most efficiently transported in free form by buoyancy forces [*McAuliffe*, 1980; *Hovland and Judd*, 1988]. Since the rise of free gas is associated with an expansion of the gas due to an adiabatic decrease in pressure, buoyancy forces are increased and the upflow is accelerated.

In contrast, upward flow of pore water at hydrostatic pressures may only occur if both a migration path and an adequate hydrostatic head of water are given. As an example, discharge of pore water was inferred from geochemical pore-fluid analyses above an 1.4 Ma old sedimented basement ridge on the Juan de Fuca Ridge flank [*Wheat and Mottl*, 1994; *Shipboard Scientific Party*, 1997b]. The maximum pore-fluid upwelling speed was estimated to 2 mm/a which is ten times higher than expected for upflow due to normal compaction [*Wheat and Mottl*, 1994]. The sediments above the ridge lack the compaction gradients observed east and west of the ridge [*Shipboard Scientific Party*, 1997b], suggesting that zones of higher porosity and permeability are present, which may act as a migration path (Chapter 4). The generation of a hydrostatic head on the other hand may be associated with basement topography [e.g. Lowell, 1980; Hartline and Lister, 1981; Fisher et al., 1990; Wang et al., 1997]. A model, which was outlined by Davis and Becker [1994] to explain driving forces for fluid vents at Middle Valley (Juan de Fuca Ridge), and which was later referred to as "choked chimney effect" by Davis and Fisher [1994], is now applied to the situation at the buried basement ridge (Figure 3.1a).

Since fluid density is dependent on temperature, different thermal gradients in the vicinity of the ridge will cause lateral differences in hydrostatic pressure. In the permeable upper basement, these pressure differences can act on fluid and cause fluid flow towards the ridge when the resistance to flow is sufficiently low. As a result, the fluid pressure beneath the ridge is shifted to a higher value, and fluid upflow from the ridge into the overlying sediments may occur. It can be expected that upflow will persist as long as a permeable connection to the seafloor exists.

Three implicit assumptions are most likely valid in this area: (1) Sediment thickness is variable above rough basement, because turbiditic sedimentation has caused relatively flat depositional planes. (2) The thermal gradient within the sediment column is linear [*Wheat and Mottl*, 1994, *Shipboard Scientific Party*, 1997a]. (3) At least the upper part of the permeable basement is isothermal as inferred from heatflow data, which are positively correlated with basement topography [e.g. Davis et al., 1992]. This can be explained by vigorous fluid circulation within the permeable parts of the crust [e.g. *Fisher and Becker*, 1995; *Davis et al.*, 1997a; *Davis et al.*, 1997b].

At two locations at a reference level beneath a trough and a ridge, respectively (Figure 3.1a), the difference in hydrostatic pressure is calculated using empirical values for sea water density [*Davis and Becker*, 1994]. For a basement elevation of 200 m, a sediment thickness above the ridge of 40 m, and a temperature difference between basement and seafloor of about 40°C [*Shipboard Scientific Party*, 1997a] the model predicts a pressure difference of 9 kPa (Figure 3.1b). The value would be three times higher for an elevation of 300 m and a temperature difference of 63°C, as observed at ODP Leg 168 Sites 1026 and 1027 further to the East. This is consistent with an estimation of *Fisher et al.* [1997] as well as with results of numerical models of hydrothermal circulation. In general, the calculated pressure difference depends linearly on basement elevation and





b

Figure 3.1 (a) Model of an 1.4 Ma old buried basement ridge on the eastern Juan de Fuca Ridge flank. The topography of highly permeable isothermal basement is associated with the generation of the hydrostatic head of water required for an upflow of pore water through relatively low permeable sediments. (b) In general, the calculated pressure difference between a basement high and a basement trough at a reference level depends linearly on basement elevation and quadratically on basement temperature, assuming that the pressure dependence of fluid density can be neglected.

quadratically on basement temperature (Figure 3.1b), assuming that the pressure dependence of fluid density can be neglected [*Furbish*, 1997].

As a result, a basement overpressure of only 9 kPa appears to maintain the observed pore fluid upflow at a rate of 2 mm/a. However, 9 kPa are most likely not sufficient to compensate for the lithostatic load of much more than one meter of sediment and can not prevent normal compaction as it is observed above the ridge. An alternative explanation for the abnormal porosity profiles may be given by differences in lithology or by a possible alteration of sediments as a consequence of fluid flow (compare to section 4.9).

However, the presented model is simple and may not be valid in general, even if hydrostatic conditions and an isothermal basement can be assumed. For example, the thermal gradient may be affected by conductive refraction which depends on thermal conductivity and basement and seafloor topography. Furthermore, realistic values for the overpressure at basement highs are probably smaller than model values, because even in highly permeable media some resistance to flow is present. Therefore, estimated values of pressure shifts are only considered to be upper limits.

3.4 Compaction flow and generation of pore pressures in excess of hydrostatic

Compaction of sediments and the generation of pore pressures in excess of hydrostatic are important processes which are often related to fluid upflow within thick sediment sections. When sediments compact under the weight of the accumulating overburden, porosity decreases with depth and the pore fluid is redistributed. Under normal conditions, the vertically ascending pore fluid does not reach the sediment surface since the ongoing sediment deposition moves the seafloor up at a faster rate than the pore fluid moves upward relative to the solid grains [*Einsele*, 1977; *Hutchison*, 1985; *Hyndman and Davis*, 1992; *Carson and Screaton*, 1998]. Thus, compaction driven upward fluid migration with respect to the seafloor requires special circumstances such as the tectonic thickening and additional strain-induced consolidation at some convergent margins [e.g. *Hyndman and Davis*, 1992].

Fluid upflow through compacting sediments may also occur when high accumulation rates combine with low hydraulic conductivity, and sediments become underconsolidated. If pore fluids are not expelled fast enough to accommodate the porosity loss and to allow the rock framework to take up the continuously increasing weight of the sediments, the fluid begins to carry part of the overburden and pore pressures in excess of hydrostatic are generated [e.g. *Bredehoeft and Hanshaw*, 1968; *Smith*, 1970; *Bethke and Corbet*, 1988]. Excess pressures, which are also known as overpressures or geopressures depending on the given pore pressure gradient [*Harrison and Summa*, 1991; *Hart et al.*, 1995], may further be increased by processes like pore space reduction due to deformation or the precipitation of minerals, aquathermal pressuring due to a temperature dependent expansion of pore water relative to the sediment matrix [*Barker*, 1972; *Sharp*, 1983], and mineral transformations such as montmorillonite to illite, which may decrease permeability and increase the volume of fluid [*Powers*, 1967; *Harrison and Summa*, 1991]. Since elevated pore pressures act to resist further compaction, some kind of transient fluid expulsion must occur to achieve an equilibrium porosity profile.

At the northern Cascadia accretionary prism, tectonic loading rather than tectonic compression is assumed to be the most important factor controlling excess pore pressure generation and fluid migration [Hyndman and Davis, 1992; Moran et al., 1995]. Approximately 3.5 km of sediment is almost completely scraped off the downgoing Juan de Fuca plate, and the thickness of the sediment wedge doubles over a distance of only 10-20 km from the deformation front [e.g. Davis and Hyndman, 1989; Hyndman and Davis, 1992]. Consequently, porosity decreases rapidly with depth and large volumes of fluid are expelled [e.g. Davis et al., 1990]. The dewatering was assumed to be mainly diffuse and hosted by intergranular and small-scale fracture permeability [Hyndman and Davis, 1992; Kastner et al., 1995]. However, the diffuse seepage appears to be superimposed by different types of confined fluid release, which are most likely related to local episodic fault displacement (Chapter 6).

3.5 Faults in convergent settings

Tectonic deformation may generate or augment elevated fluid pressures, but may also create faults, which act as potential pathways for a subsequent fluid release. Enhanced permeability may be associated with ongoing vertical displacement along a fault. However, measured flow rates and mass balance calculations indicate that focused fluid flow along faults at convergent margins is rather episodic and transient than continuous [e.g. *Roberts et al.*, 1996]. This implies that fault zone permeability is actively reduced by mechanisms such as a collapse of the fracture network due to the drop in fluid pressure, the precipitation of mineral cements along the fault zone, and sediment compaction in and around the fault zone [e.g. *Roberts et al.*, 1996]. Thus, fault zones may change from relative permeable conduits to low permeable barriers and vice versa.

At the northern Cascadia accretionary prism, major thrust faults are absent and focused fluid upflow is mainly related to local processes. Local tectonism creates topographic depressions, which are subsequently filled with bedded hemipelagic deposits. Growth faults are active over long periods of time and reflect the development of those sedimentary basins. However, they would not allow for focused fluid flow due to their limited vertical extent of faulting during the intermittent tectonic activity. It is more likely, that episodic fault displacement along normal faults, which affects the complete sediment column between the seafloor and the BSR, triggers single events of fluid expulsion, followed by periods, when the basin deposits act as an efficient seal. One consequence of the transient fluid expulsion may be the cementation of seafloor sediment layers due to carbonate precipitation (Chapter 6).

3.6 Formation of mineral deposits

If a liquid becomes saturated or supersaturated with solutes, the balance between the solvent and the solutes is governed by temperature, pressure, and solute concentration. A decrease in temperature or pressure, or an increase in concentration, generally leads to the precipitation of excess solutes [e.g. *Hovland and Judd*, 1988]. As an example, carbonate precipitation is often directly related to focused pore water expulsion (cold seeps) and the exposure of dissolved hydrocarbons to the low-temperature and low-pressure oxidizing conditions near the seafloor [e.g. *Ritger et al.*, 1987; *Kulm and Suess*, 1990; *Suess et al.*, 1998]. At the sediment surface, carbonate cements may form slabs, edifices, or chimneys [*Kulm and Suess*, 1990] as well as crusts, which may extend for distances of more than 250 m [*Carson et al.*, 1994].

In the vicinity of subseafloor magmatic heat sources, different mineral phases may form from hydrothermal solutions (hot seeps) as a product of chemical and thermal exchange processes between lithosphere and ocean. The precipitaton of metallic oxides, hydroxides, and silicates, which is inferred to reduce permeability and constrain fluid circulation, is due to hydrothermal activity at temperatures below 200°C [e.g. Stein and Stein, 1994; Lowell et al., 1995], whereas ores (sulfides) are formed at discharge temperatures of 200-400°C [e.g. Lowell, 1995].

However, the temperature of the 1.4 Ma old buried basement ridge on the Juan de Fuca Ridge flank is about 40°C (Figure 3.1), and hydrothermally driven fluid upflow at a rate of 2 mm/a can only be detected by chemical pore fluid analyses. There is presently no evidence for mineral precipitation within the sediments or a positive feedback between hydrothermal discharge and sediment properties at the location of the ridge [*Giambalvo et al.*, 1998].

Chapter 4: Seismic reflectivity anomalies in sediments at the eastern flank of the Juan de Fuca Ridge: Evidence for fluid migration?

Lars Zühlsdorff, Volkhard Spieß, Christian Hübscher, and Monika Breitzke, Journal of Geophysical Research, 104, 15,351-15,364, 1999.

4.1 Abstract

A high-resolution seismic survey was carried out in the vicinity of Ocean Drilling Program (ODP) Leg 168 drill sites at the eastern flank of the Juan de Fuca Ridge. Three seismic systems with different source frequencies up to 4000 Hz were used simultaneously along each seismic line. The data sets were combined to provide the best possible resolution at any given depth level. An integrated interpretation of narrow zones of low reflection amplitudes is presented, which are related to basement highs but are independent of seafloor topography. The effect is most pronounced above a buried basement ridge in the vicinity of ODP Sites 1030 and 1031, where higher porosity values are present. Higher porosities can reduce seismic impedance contrasts and may therefore cause the observed low reflection amplitudes. In addition, Biot-Stoll's [Biot, 1956a, b; Stoll, 1989] model suggests that attenuation coefficients are porosity dependent and reflection amplitudes are further decreased at higher seismic frequencies when porosity increases. Since zones of higher porosities are potential pathways for fluids, they may be associated with hydrothermally driven fluid discharge, which was previously predicted above the buried basement ridge. It is therefore proposed that the observed seismic amplitude reduction in the vicinity of ODP Sites 1030 and 1031 indicates zones of upward fluid migration.

4.2 Introduction

Thermal contraction opens cracks within cooling rocks after young crust is formed along oceanic spreading centers [e.g., *Hartline and Lister*, 1981]. Fluid enters these cracks and supports further cooling and cracking. A permeable crust made up of interconnected pores and fractures is left behind. From observations of hydrothermal activity at mid-ocean ridges, it is known that large-scale fluid flow occurs within the young basement rocks [e.g., *Lister*, 1972; *Fisher et al.*, 1994; *Davis et al.*, 1997a; *Wang* et al., 1997].

Until the basement rocks are covered by sediments or sealed by chemical alteration, about 50 ocean volumes circulate through the upper crust, accounting for the largest portion of heat loss in young oceanic lithosphere [*Stein and Stein*, 1994; *Davis and Chapman*, 1996]. Direct exchange of ocean water will later be diminished by a growing sediment cover of lower permeability [e.g., *Lister*, 1972; *Snelgrove and Forster*, 1996]. However, even after a sediment seal isolates crustal fluid motions from the ocean, convective processes within the crust may still transport large quantities of fluid and heat [*Shipboard Scientific Party*, 1997b].

Advective fluid flow within the sediments is controlled by hydraulic and physical properties of the crust and the overlying sediment cover. A significant proportion of crustal cooling is still nonconductive, and as much as 70% of the advective heat flux through oceanic crust was estimated to occur on ridge flanks older than 1 Ma [*Stein and Stein*, 1994]. Fluid upflow velocities of the order of millimeters per year through the sediments can be inferred from geochemical observations. These velocities, however, are too low to be detected thermally using short heat flow probes [*Wheat and Mottl*, 1994].

Factors such as heat distribution, permeability structure, sediment thickness, completeness of sediment cover, and basement topography control the extent and effects of off-axis hydrothermal fluid migration [e.g., *Fisher et al.*, 1994]. Relations between these factors are complex in most marine settings. Fluid flow, on the other hand, influences the physical and chemical properties of crust and sediments in different ways. It accounts for transport of dissolved metals and sulfur, accelerates alteration and mechanical consolidation, and affects the heat budget of marine environments [e.g., *Davis and Becker*, 1994]. Exchange processes at oceanic spreading centers therefore have a general impact on ocean chemistry. However, detailed knowledge about the geological and physical parameters of hydrothermal systems is limited, and the nature and scale of fluid and heat transport at ridge flanks is still poorly understood [e.g., *Davis and Becker*, 1994].

The Cascadia Basin (Figure 4.1a) and the adjacent Juan de Fuca (JDF) Ridge are appropriate to study fluid migration systems in greater detail. Massive sediment input


Figure 4.1 (a) Ocean Drilling Program (ODP) Leg 168 study area (modified from *Shipboard Scientific Party* [1997a]), (b) location of ODP Leg 168 drill sites (modified from *Shipboard Scientific Party* [1997a]), and (c) location of ODP Leg 168 sites and seismic lines presented in this paper (bold segments). Seafloor bathymetry was provided by the Hydrosweep multibeam swath sounder (contour levels 10 m).

from the continent provides a low-permeable sediment cover above young and permeable upper crust [e.g., *Davis et al.*, 1992a]. Within a relatively small area, three different types of young crust can be studied (Figure 4.1b): a transition zone between sediment-free and sediment-covered basement near the JDF ridge crest, an adjacent area of flat basement beneath a homogeneous sediment cover, and sites of outcropping basement further to the east, providing direct hydraulic connections between basement and ocean [*Davis et al.*, 1992a]. Ocean Drilling Program (ODP) Leg 168 drill holes penetrate the sediments and the upper basement in all three areas. The project was targeted to elucidate the fundamental physics and fluid chemistry of ridge-flank hydrothermal circulation and the consequent alteration of the upper igneous crust and sediments that host the flow [*Shipboard Scientific Party*, 1997a].

Shortly after ODP Leg 168 was finished, very high-resolution multichannel seismic data were collected during R/V Sonne Cruise SO111. Lines were crossing most of the ODP Leg 168 drill holes. The main study area was located near the transition zone between sediment-free and sediment-covered parts of the young oceanic crust in the vicinity of ODP Sites 1030 and 1031 (Figures 4.1b and 4.1c). The objective was to identify fluid upflow zones through the sediments as were previously predicted by *Wheat and Mottl* [1994], to investigate the shape and extent of these zones, and to image the associated sedimentary structures.

In this paper the results of the high-resolution seismic survey are presented and compared to previously collected data and the preliminary findings of ODP Leg 168.

4.3 Study area

The sediments of the Cascadia Basin consist of sand-rich and silt-rich Pleistocene turbidites up to 1 m thick interbedded with hemipelagic mud typically tens of centimeters thick [Shipboard Scientific Party, 1997a]. Sediment thickness varies between zero at the ridge crest and more than 3 km near the base of the continental slope in the east [Davis and Hyndman, 1989], generally increasing stepwise between a number of relatively flat basement plateaus. At smaller scales, however, sediment thickness is highly variable because of local relief of the basement [Davis and Hyndman, 1989; Davis et al., 1992a], where buried ridges strike parallel to the JDF

ridge axis in a northeastern direction. They were produced by normal faulting and variations in volcanic supply at the time the crust was created [e.g., *Kappel and Ryan*, 1986].

The first buried basement ridge (ODP Leg 168 Sites 1030 and 1031), which is located about 40 km east of the JDF ridge crest on 1.4 Ma old crust (Figures 4.1b and 4.1c), was the main target for the high-resolution seismic survey. Along this ridge, sediment thickness varies between 20 and 150 m. At sites of minimum sediment thickness, fluid discharge was inferred from geochemical pore-fluid analyses [*Wheat and Mottl*, 1994]. Pore-fluid profiles later provided additional evidence of upward fluid flow at ODP Sites 1030 and 1031 [*Shipboard Scientific Party*, 1997b]. The corresponding porosity profiles are relatively constant with depth and lack the compaction gradients observed near the surface of all adjacent ODP sites. Sediments from Holes 1030 and 1031 are further characterized by a relatively high carbonate content and very little sand. However, whether the observed porosity profiles reflect differences in lithology or active porewater upwelling is uncertain at present [*Shipboard Scientific Party*, 1997b].

4.4 Seismic equipment and data processing

The following systems were used for the survey.

4.4.1 Hydroacoustic systems

The narrow-beam Parasound echosounder (4 kHz, opening angle 4°) provides detailed acoustic images of the upper 50-100 m of the sediment column at very high-resolution. With this system, seafloor reflections can be recorded for inclinations up to 2°. Additionally, the Hydrosweep multibeam swath sounder maps the seafloor bathymetry and shows small-scale features at the seafloor.

4.4.2 Multichannel seismics

All components of the multichannel seismic system were designed to optimize lateral and vertical resolution:

1. The data acquisition unit (Bison Spectra) was specially developed to work at high shot and sampling rates. At the Cascadia Basin, shots were recorded every 10 s at a sampling rate of $125 \,\mu$ s.

2. The streamer (300 m, Syntron Inc.) allowed for separately programmable

hydrophone subgroups, which were optimized for the given target and resolution. In the Cascadia Basin, 24 groups at a length of 6.25 m were used. Four remotely controlled birds kept the streamer depth within a range of 1 m, and magnetic compass readings allowed for the determination of the position of each streamer group relative to the ship's course.

3. Two seismic sources, a Sodera S 15 water gun (0.16 L) and a GI-Gun with reduced volumes (2 x 0.4 L), provided frequency ranges of 200-1600 Hz and 100-400 Hz, respectively. Signal penetration was dependent on the type of sediment and reached 150 m for the water gun and more than 1 km for the GI-Gun.

All systems were used simultaneously along each seismic line (seismic sources in an alternating mode), improving the accuracy of navigation and minimizing required ship time.

Properties of the different data sets as well as standard seismic processing steps are summarized in Table 4.1. Custom software was developed to apply a static correction for the streamer-towing depth and to determine geographic positions of common depth points (CDPs), accounting for streamer drift, variations in ship velocity, and irregularities in trigger time. A spacing of 20 m between successive CDP bins was chosen, which corresponds to a fold of nine traces for each bin. As a consequence, migration of high-frequency water gun data is hampered by aliasing when seafloor inclination exceeds 1°. However, when a smaller bin spacing is used, the S/N ratio also decreases significantly. An appropriate comparison between the different data sets in the vicinity of the sedimented ridge is therefore done for stacked sections only.

	Gl-Gun	water gun	Parasound
Bandwidth, kHz	0.1-0.4	0.2-1.6	2.0-6.0
Radius of Fresnel zone*, m	88 (at 250 Hz)	44 (at 1000 Hz)	22 (at 4000 Hz)
Trace spacing, m	20 (bin spacing)	20 (bin spacing)	5 (average trace spacing)
Fold	9	9	1
Processing	filter, CDP-stack, migration	filter, CDP-stack	filter
Processing	hiter, CDP-stack, migration	filter, CDP-stack	

Table 4.1 Properties of data sets

*Calculated for a depth of 2600 m.

All data were carefully processed to preserve the complete amplitude information. Only spherical divergence was corrected by multiplying with travel time and assuming a constant velocity for each trace. For seismic velocities estimated from the data, an error of about 10% could be expected because of the short streamer length. However, observed velocity variations appear to be small, and a constant sediment velocity of 1500 m/s is in agreement with velocity profiles from the ODP sites [Shipboard Scientific Party, 1997b, c].

4.5 Data sets

Bold lines in Figure 4.1c indicate the location of seismic lines presented in this paper. All lines run perpendicular to the known basement ridges, except for Line GeoB 96-074, which is oriented parallel to the first sedimented ridge across ODP Leg 168 Sites 1030 and 1031. Depth scale and vertical exaggeration are the same for Figures 4.2 to 4.5 and were calculated assuming a constant velocity of 1500 m/s. For display, all seismic sections were muted manually.

Line GeoB 96-081 (Figure 4.2) starts about 25 km east of the JDF ridge crest and ends 3 km west of the first sedimented basement ridge. Sediment thickness is less than 150 m. GI-Gun data at the lowest resolution of the three data sets reach sufficient signal penetration to reveal the morphology of rough oceanic basement (migrated section, Figure 4.2a). The stacked section (Figure 4.2b) can then be compared to the stacked water gun data.

The penetration of the water gun signal is lower (same line, Fig 2c), but sediment structures are imaged in greater detail. The most outstanding features are lateral amplitude changes at some locations with a horizontal extent of several tens of meters (e.g., at CDPs 400 and 460).

The corresponding Parasound data in Figure 4.2d show the same zones of low reflectivity, which are more pronounced at higher frequency and resolution. Depth penetration, however, is limited to 50 m.

If the corresponding data sets are jointly analyzed, two additional features can be observed, which could not be derived from only one single data set. (1) Zones of decreased reflection amplitudes at CDPs 400 and 460 in the water gun record can be connected to areas of basement highs in the GI-Gun record. (2) The amplitude decrease is observable at all different source frequencies but appears strongest at 4 kHz (Parasound) and weakest at 100-400 Hz (GI-Gun).



460 in the water gun record (200reflection amplitudes at common decrease is visible at all different and weakest at 100-400 Hz (GI-Figure 4.2 Line GeoB 96-081: assuming a constant velocity of source frequencies but appears Figures 4.2-4.5 were calculated GI-Gun record. The amplitude strongest at 4 kHz (Parasound) (a) GI-Gun (migrated), (b) GI-1500 m/s. Zones of decreased 1600 Hz) can be connected to areas of basement highs in the depth points (CDPs) 400 and (stacked), and (d) Parasound. For location, see Figure 4.1c. Vertical exaggeration (VE) for Gun (stacked), (c) water gun Gun).

Zones of low reflection amplitudes above basement highs are observed in all lines across the first sedimented ridge. They can be traced from the basement to the seafloor and are often present where sediment thickness decreases to less than 50 m. Lines GeoB 96-101 (Figure 4.3) and GeoB 96-104 (Figure 4.4) show the most pronounced amplitude effects. Both lines are located in the vicinity of ODP Leg 168 Site 1030 and are typical for most of the lines across the ridge (Figure 4.1c). To evaluate the degree of amplitude decrease, root-mean-square (rms) amplitudes were calculated for a time window of 20 ms length beneath the seafloor reflection (Figures 4.3e and 4.4e). It is sufficiently long to include a large number of amplitude values, but it also ensures that basement amplitudes are not sampled by mistake. The time window follows the seafloor topography to account for lateral changes in signal attenuation within the surface sediments. For display, the values are further smoothed along the horizontal axis by a median filter of 200 m length and are normalized to the average rms amplitude of each data set. Above the first buried ridge, GI-Gun amplitudes drop by 40-45% (solid lines), water gun amplitudes drop by 50-55% (dashed lines), and Parasound amplitudes drop by 65% (dotted lines).

Line GeoB 96-074 (Figure 4.5) along the axis of the first buried ridge passes ODP Sites 1030 and 1031 at distances of 70 and 135 m, respectively. Between CDPs 130 and 230, sediment thickness is below 50 m, and wide intervals of low reflection amplitudes are present. This observation is consistent with amplitude anomalies at the crossing lines GeoB 96-101 and GeoB 96-104.

To infer the areal variation of reflection amplitudes, rms amplitudes as displayed in Figures 4.3e and 4.4e were calculated along all lines across the first sedimented ridge (Figure 4.1c). After gridding the smoothed data with continuous curvature splines [*Smith and Wessel*, 1990], an additional two-dimensional median filter of 300 m diameter was applied for further smoothing. Near-surface reflectivity was then mapped for GI-Gun (Figure 4.6a), water gun (Figure 4.6b), and Parasound (Figure 4.6c) data. Contour levels give the deviation from the average rms value of each individual data set, and dark colors indicate low reflection amplitudes. As a result, distinct amplitude anomalies can consistently be identified in all three data sets. However, the observed amplitude decrease appears to be more pronounced at higher frequencies and is



Figure 4.3 Line GeoB 96-101 across the first sedimented ridge (for location see Figure 4.1c); the intersection with crossing line GeoB 96-074 is indicated: (a) GI-Gun (migrated), (b) GI-Gun (stacked), (c) water gun (stacked), (d) Parasound, and (e) decrease of rms amplitudes (20 ms time window) normalized to an average value and smoothed by a median filter of 200 m length along the line. The time window follows the seafloor topography to account for lateral changes in signal attenuation. Above the first buried ridge, GI-Gun amplitudes drop by 45% (solid line), water gun amplitudes drop by 55% (dashed line), and Parasound amplitudes drop by 65% (dotted line).



Figure 4.4 Line GeoB 96-104 across the first sedimented ridge (for location see Figure 4.1c); ODP Site 1030 and the intersection with crossing line GeoB 96-074 are indicated: (a) GI-Gun (migrated), (b) GI-Gun (stacked), (c) water gun (stacked), (d) Parasound, and (e) decrease of rms amplitudes as described for Figure 4.3. Above the first buried ridge, GI-Gun amplitudes drop by 40% (solid line), water gun amplitudes drop by 50% (dashed line), and Parasound amplitudes drop by 65% (dotted line).





strongest for Parasound data.

The location of Parasound reflectivity anomalies is then compared to both seafloor bathymetry and basement topography (Figure 4.7). Basement topography was provided by manually picking basement reflections from migrated seismic lines and using a constant velocity of 1500 m/s to convert reflection time into depth. Seafloor reflections were also picked from all lines to estimate sediment thickness from the time difference between seafloor reflection and basement reflection. In Figure 4.7 the Parasound reflectivity map (Figure 4.6c) is overlaid by contour levels of water depth (Figure 4.7a), basement depth (Figure 4.7b), and sediment thickness (Figure 4.7c). Seafloor reflectivity is apparently not correlated with seafloor bathymetry since distinct areas of low reflection amplitudes occur at different depth levels and do not seem to be related to a certain range of seafloor inclinations. In contrast, amplitude anomalies appear to be associated with pronounced basement topography and low reflection amplitudes is observed along the main basement ridge, which is roughly parallel to Line GeoB96-074, and at a more isolated basement high to the west.

4.6 Causes for lateral amplitude changes

The reduction of rms amplitudes above the first buried ridge is significant for all data sets at different source frequency ranges (Figures 4.3e, 4.4e, and 4.6). The similarity of results for GI-Gun, water gun, and Parasound data suggests that rms amplitudes are basically independent of the seismic source signal, which is characterized by frequency, footprint size, and signal generation method. However, a slight frequency dependence can be observed as the pronounced amplitude decrease above the first sedimented ridge is more enhanced at higher source frequencies.

Possible explanations for lateral amplitude changes are discussed below and include (1) artifacts due to data acquisition and processing, (2) effects of topography, (3) scattering, and (4) lateral changes of sediment physical properties. The effect of free gas on seismic reflection amplitudes is not discussed since methane concentrations in the sediments from ODP Sites 1030 and 1031 are close to background values and volatile hydrocarbons other than methane are absent [*Shipboard Scientific Party*, 1997b].



Figure 4.6 Areal distribution of reflection amplitude anomalies for (a) GI-Gun, (b) water gun, and (c) Parasound data. Rms values as described for Figure 4.3 were further smoothed in two dimensions by a median filter of 300 m diameter. Contour levels give the deviation from the average rms value of each data set with dark colors indicating low reflection amplitudes.



Figure 4.7 Comparison of Parasound reflectivity anomalies (Figure 4.6c) with (a) seafloor topography (10 m contour levels, compare to Figure 4.1), (b) basement topography (20 m contour levels), and (c) sediment thickness (20 m contour levels). Dark shading indicates low reflection amplitudes.

1. During data acquisition, lateral variation of source energy was small. A potential residual effect was minimized by stacking nine traces of different shots for each CDP bin. Further, all data were processed carefully to preserve the complete amplitude information. Amplitude anomalies in seismic data can be compared to similar effects in Parasound data (Figures 4.3e and 4.4e) which are heave, pitch, and roll compensated and are completely independent with regard to signal generation and timing. It is therefore assumed that lateral amplitude changes are not related to processing artifacts.

2. Topography may be a possible explanation, as the most pronounced zones of low reflection amplitudes are observed in the vicinity of a buried basement ridge. However, since Parasound reflections were still recorded, the seafloor inclination must be less than 2° . Zones of low reflectivity are not linked to a specific location with respect to seafloor inclination or distance from the sedimented ridge crest and occur at the ridge flanks as well as on flat seafloor (Figure 4.7a). In areas of moderate topography, amplitude anomalies can further be traced down to flat reflectors in greater depth (Figure 4.2). Since the velocity contrast between seawater and surface sediment is very small, no refraction occurs at the seafloor or at deeper reflectors, and lateral amplitude variations at flat reflectors cannot be explained by an effect of geometry. In addition, Figure 4.7 suggests that reduced reflection amplitudes are correlated with basement elevation rather than with seafloor bathymetry. It is therefore very unlikely that topography has a significant effect on the decrease of reflection amplitudes.

3. Reduced reflection amplitudes due to scattering are also unlikely because scattering features must show a wide range of sizes to account for similar observations as made for the frequency range between 200 and 4000 Hz. ODP Leg 168 provides no evidence for any extreme lithologic or structural inhomogeneities at scales of some centimeters to some meters due to intense deformation or diagenesis.

4. The most likely reason for lateral amplitude variations is a lateral change in sediment physical properties. Support for this hypothesis is given by the preliminary findings at ODP Sites 1030 and 1031, which are located in the vicinity of zones of reduced reflection amplitudes (Figure 4.6). Cores from these sites contain a relatively small proportion of sand and do not reveal compaction-related reduced porosities with depth. Also, the lowest bulk densities compared to all other locations along the buried

basement transect of Leg 168 are observed at comparable depth [Shipboard Scientific Party, 1997b]. Since the bulk density of saturated sediments is mainly controlled by porosity, lateral porosity changes may be the main reason for the observed amplitude anomalies.

4.7 Impact of porosity on seismic reflection amplitudes

A simple approach is used to demonstrate how a lateral increase in porosity may cause a lateral decrease in seismic reflection amplitudes. This implies that either reflection coefficients are decreased or seismic attenuation is increased. Below, a rough estimation based on average values shows that the observed variation of rms amplitudes (Figures 4.3e and 4.4e) can be explained by a combination of both effects.

The normal incidence reflection coefficient for plane waves and between media of densities ρ_1 and ρ_2 and velocities v_1 and v_2 is given by $R=(v_2\rho_2-v_1\rho_1)/(v_2\rho_2+v_1\rho_1)$. Wet bulk densities ρ_1 and ρ_2 are weighted means of fluid density ρ_f and grain density ρ_g . They can be discussed in terms of porosity ϕ as the main controlling factor. In the case of constant fluid density, this relationship is given by $\rho_{1,2}=\rho_f\phi+\rho_{g1,2}(1-\phi)$. Variations of grain density are considered to be of minor importance, and changes of fluid density due to dissolution or precipitation can also be neglected in deep sea surface sediments. However, significant changes of porosity will strongly affect wet bulk density.

In contrast, the dependence of velocity on porosity is weaker since typical relative variations of velocity in surface sediments are small (< 5%) compared to relative density changes (> 10-20%), especially when porosities are above 50% [e.g., Breitzke, 1999; Weber et al., 1997]. Hence, in fine grained marine surface sediments, seismic impedance contrasts are usually dominated by density contrasts. Above the first buried ridge, fine-grained sediments are abundant [Shipboard Scientific Party, 1997b], and sandy turbidites are absent because of the higher elevation. Also, reflection coefficients of seismic data show no indication for massive turbidites. For simplification, the reflection coefficient R is calculated only for porosity-dependent variations in density, whereas velocity is assumed to be constant.

For constant grain densities in both media, the reflection coefficient (or the contrast of wet bulk densities ρ_1 and ρ_2) decreases when porosity increases. Figure 4.8 shows

relative changes of R with respect to a reference porosity of 65%. According to this example, a lateral increase in porosity from 65 to 82% would cause a decrease in R of 40%, which is consistent with the maximum decrease of GI-Gun rms amplitudes (Figures 4.3e and 4.4e).

For distinct turbiditic layers this estimate may be too simple because, in this case, velocity also contributes to the reduction of the reflection coefficient when porosity increases. In terrigeneous sediments, however, velocity changes are normally associated with a much more pronounced change in density [e.g., *Weber et al.*, 1997]. Therefore the main decrease of the reflection coefficient will still be due to a decrease of the density contrast. It is expected that the porosity increase needed to explain the observed amplitude effects is overestimated by 10-20% when velocities in distinct turbidite layers contribute to R.

As demonstrated above, lateral changes of reflection coefficients can account for a major portion of the observed amplitude effects. In addition, a possible impact of porosity on seismic attenuation was investigated using Biot-Stoll's [Biot, 1956a, b; Stoll, 1989] model to compare results from different seismic sources. Seismic velocities and attenuation coefficients were calculated for different porosities and typical frame and grain parameters of terrigeneous sediments. Since porosity is expected to affect permeability, which has significant impact on seismic attenuation, porosity-dependent permeability k was calculated using Kozeny-Carman's [Carman, 1956] equation. This equation depends also on mean grain size d and a dimensionless constant k and is given by $\kappa = (d^2 \phi^3)/(36k(1-\phi)^2)$. The constant k was set to 5, which is a standard value for sediments with higher porosities [Hovem and Ingram, 1979; Hovem, 1980; Breitzke et al., 1996]. Permeabilities were calculated for constant grain sizes of 6 µm and porosities between 40 and 80% (Table 4.2). Values are consistent with results of laboratory measurements for silt-rich surface sediments of the Cascadia Basin [Snelgrove, 1994; Snelgrove and Forster, 1996]. For the same porosity range the model yields seismic velocities between 1500 and 1550 m/s, that is, a variation of only 3% for the seismic frequencies used. In contrast, porosity-dependent permeability has strong impact on frequency-dependent attenuation (Figure 4.9). Attenuation coefficients generally increase with increasing frequency, and the attenuation effect is more pronounced at



Figure 4.8 Decrease of the plane wave normal incident reflection coefficient with increasing porosity, assuming constant velocity and grain densities. Values are plotted with respect to a reference porosity of 65%. For example, an increase in porosity from 65 to 82% causes a decrease in R of 40%.



Figure 4.9 Porosity-dependent attenuation coefficients for grain sizes of 6 μ m, a constant factor of 5 for Kozeny-Carman's [*Carman*, 1956] equation, and typical parameters for terrigeneous sediments. Attenuation coefficients generally increase with increasing frequency, but the attenuation effect is more pronounced at higher porosities.

higher porosities. According to Figure 4.9, signals at typical frequencies of Parasound and water gun (Table 4.1) are more strongly attenuated at high porosities than at low porosities, whereas this effect is less pronounced at typical GI-Gun frequencies.

Porosity, %	Permeability, m ²	Grain Size, µm
40	3.56x10 ⁻¹⁴	6
50	1.00x 10 ⁻¹³	6
60	2.70x10 ⁻¹³	6
70	7.62x10 ⁻¹³	6
80	2.56x10 ⁻¹²	6

Table 4.2 Calculated permeability values for different porosities using Kozeny-Carman's equation with a dimensionless constant of 5. For reference, laboratory measurements indicate a permeability of about 10⁻¹³ m² for silt-rich sediment near the seafloor [*Snelgrove*, 1994; *Snelgrove and Forster*, 1996].

Frequency-dependent attenuation coefficients and velocities provided by Biot-Stoll's model were used to determine amplitude and two-way time (TWT) of a model wavelet. Waveforms were calculated with respect to a fixed sediment thickness of 40 m and are displayed for medium porosities of 40-80% and wavelet center frequencies of 400 (Figure 4.10a), 1000 (Figure 4.10b), and 4000 Hz (Figure 4.10c). As a result, porosity-dependent attenuation can be neglected for signal frequencies up to 400 Hz (Figure 4.10a). At typical frequencies for water gun and Parasound, however, attenuation increases significantly when porosity increases (Figures 4.10b and 4.10c). According to this model, the porosity-dependent portion of seismic attenuation would decrease water gun amplitudes by 7% with respect to GI-Gun amplitudes when porosity increases from 65 to 82%. Parasound amplitudes would even be 22% lower than GI-Gun amplitudes at the same porosity change.

Although this result is in reasonable agreement with observations in Figures 4.3e and 4.4e, it provides only a rough estimate of the expected amplitude decrease in measured data. For a homogeneous medium a sediment thickness of 40 m is needed to explain the observed variation of rms amplitudes at different signal frequencies. In contrast, the effect is less pronounced near the sediment surface. However, a vertical variation of grain size is observed within the sediment column, and larger grains of sandy or silty turbiditic layers would increase attenuation effects. In addition, attenuation coefficients may vary strongly with the dimensionless constant k. Independent of model parameters, a porosity-dependent attenuation effect is always present through the associated change



Figure 4.10 Amplitude and two-way time (TWT) of a model wavelet with respect to a fixed sediment thickness of 40 m and medium porosities of 40-80%. Frequency-dependent attenuation coefficients and velocities were provided by Biot-Stoll's [*Biot*, 1956a, b; *Stoll*, 1989] model. Results are presented for wavelet center frequencies of (a) 400, (b) 1000, and (c) 4000 Hz.

in permeability, and hence attenuation generally increases with porosity even if grain size is constant.

In summary, a lateral porosity increase from 65 to 82% may explain an overall decrease of reflection coefficients by 40% as observed for GI-Gun data (Figures 4.3e and 4.4e). Biot-Stoll's model suggests that the same porosity change may also explain an additional amplitude decrease for water gun and Parasound data due to porosity-dependent attenuation, whereas the porosity-dependent portion of attenuation can be neglected for typical frequencies of the GI-Gun. Therefore the proposed amplitude changes due to lateral porosity variations are in agreement with observations in all three data sets. It can further be shown that seismic velocities do not depend strongly on porosity when porosity exceeds 50% (Figure 4.10).

4.8 Upward fluid migration at the first sedimented ridge

In the previous section it was demonstrated that zones of decreased seismic reflection amplitudes can be explained by lateral porosity variations. In this section, the evidence for a relationship between high porosity values and fluid upflow at the first sedimented ridge near ODP Leg 168 Sites 1030 and 1031 is renewed.

Indications for fluid discharge at the first sedimented ridge were provided by chemical data. *Wheat and Mottl* [1994] presented near-surface pore-fluid analyses for nine cores in the vicinity of the ridge. Three cores revealed evidence suggesting fluid upflow, although data were not available from locations of maximum heat flow where the highest upflow rates may be expected. The maximum pore-fluid upwelling speed was estimated to be 2 mm/yr, which is 10 times higher than expected for upflow due to normal compaction. It was further estimated that fluid upflow may occur as long as sediment thickness is less than 160 m [*Wheat and Mottl*, 1994]. Data from ODP Leg 168 Sites 1030 and 1031 are consistent with the results of *Wheat and Mottl* [1994] since fluid upflow was inferred from pore-fluid profiles at both sites above the first buried ridge [*Shipboard Scientific Party*, 1997a].

It has long been discussed that buoyancy-driven flow may be governed by basement topography [e.g., Lowell, 1980; Hartline and Lister, 1981; Fisher et al., 1990; Wang et al., 1997], which is consistent with the observations at ODP Sites 1030 and 1031. Fluid

discharge occurring above the buried basement ridge indicates that some local fluid overpressure must be present. On the basis of hydrostatic principles and by assuming isothermal conditions in the upper crust, an upper limit of this overpressure can be estimated to be 9 kPa. Evidence for an isothermal upper basement is provided by heat flow data, which are positively correlated with basement topography [e.g., *Davis et al.*, 1992a]. This can be explained by vigorous fluid circulation within the permeable parts of the crust [e.g., *Fisher and Becker*, 1995; *Davis et al.*, 1997a, b].

A basement overpressure of only 9 kPa appears to maintain the observed pore-fluid upflow at a rate of 2 mm/yr at locations which are characterized by higher porosity. However, 9 kPa are most likely not sufficient to compensate for the lithostatic load of much more than 1 m of sediment and cannot prevent normal compaction as it is observed at ODP Sites 1030 and 1031.

4.9 Discussion

Zones of reduced seismic reflection amplitudes with a horizontal extent of several tens of meters occur at locations where previous work has shown that porosity is locally high and fluid upflow is inferred (Figures 4.3 and 4.4). Assuming that these observations are coupled, it is proposed that amplitude anomalies reflect porosity changes, which are related to fluid upflow. This upflow may be a result of the locally decreased hydraulic impedance of the sediment. As suggested by Biot-Stoll's model, zones of higher porosity are associated with higher permeability and therefore provide potential pathways for fluid migration. In contrast, when constant permeability is assumed for Biot-Stoll's model, seismic attenuation is expected to decrease with increasing porosity, which is not consistent with the observation of an attenuation-related seismic amplitude reduction above the first buried ridge.

However, interpretation of reflection amplitudes is currently limited to a comparison of relative amplitude changes and is based on average estimates of porosity and grain size. It was shown that lateral changes of porosity are, in principle, sufficient to explain the observed amplitude effects, but a more quantitative analysis is difficult. Observed porosity values of most ODP Leg 168 sites are biased toward an oversampling of mud, and average values may not represent the mean of the drilled section [*Shipboard* Scientific Party, 1997c]. Further, only two drill sites at the first buried ridge are probably not sufficient to characterize physical properties of narrow upflow zones as lateral variations over tens of meters may be significant. In fact, average porosity values between 70 and 74% at ODP Sites 1030 and 1031 are higher than those observed at all other sites but are still lower than estimated in Figure 4.8 for a reference porosity of 65% and a decrease of GI-Gun amplitudes of 40%. More drill sites would be necessary to improve the presented model and to further quantify the observed effects.

A question, which cannot be answered on the basis of seismic data, is how the presence of high-porosity zones in the vicinity of the first buried ridge can be explained. Since the estimated superhydrostatic pressure in the basement is not sufficient to compensate for the sediment load, another process is required to prevent normal compaction at ODP Sites 1030 and 1031. An alternative explanation for abnormal porosity profiles may simply be given by differences in lithology. These may be a result of vertical grain size sorting in a turbiditic particle cloud since the proportion of suspended material, which can reach elevated areas, predominantly consists of finer grains, and hemipelagic sedimentation above ridges is less diluted by turbiditic input. However, although sandy turbidites are absent above the buried ridge, analysis of the areal distribution of low reflection amplitudes and their relation to bathymetry, basement topography and sediment thickness does not support the assumption that elevation-dependent lithologic changes are the only reason for the observed highporosity zones. Distinct reflectors can be traced throughout the sedimentary basins to locations above the basement high, indicating that the upper portion of the sediment cover, which includes the major portion of sediments above the basement ridge, is uniform over wide areas with only slight changes of sedimentation rate at elevated locations (Figures 4.3a, 4.4a, and 4.5a). Further, all three seismic data sets consistently show distinct areas of low reflection amplitudes, which are located at the topographic high as well as to the west of the topographic high in a deeper situated area (Figures 4.3, 4.4, 4.6, and 4.7a). The sediments in between these two areas are characterized by average values of reflection amplitude, although they were also deposited on the basement high (Figures 4.3 and 4.4). Topography-dependent lithologic changes in a turbiditic depositional realm should provide a gradual variation of grain size and

reflectivity on larger scales and less local variability than it is observed in seismic data for some locations above the first sedimented basement ridge. However, the overall distribution of low reflectivity zones appears to be correlated with basement topography (Figure 4.7b) and sediment thickness (Figure 4.7c), suggesting that factors such as the hydraulic impedance may be important to explain the observations. It may be speculated that the high porosity is coupled to fluid upflow above basement highs. As a suggestion, fluids may alter the sediments in a way that enhances the stability of the matrix and therefore prevent normal compaction. However, there is presently no evidence for a positive feedback between hydrothermal discharge and sediment properties that could support this hypothesis [*Giambalvo et al.*, 1998].

Independent of the process that causes higher porosities at the first sedimented ridge, fluid upflow at the eastern flank of the JDF Ridge may be expected wherever permeable connections between basement highs and seafloor are present. When porosity at the first buried ridge is related to permeability, vertical zones of higher porosity are directly related to fluid upflow. Therefore it can be stated that fluid upflow in the vicinity of the first buried ridge can be detected and imaged by high-resolution seismic data with source frequencies of more than 200 Hz. Seismic surveys, combined with geological and chemical data, may therefore be a useful tool to further investigate fluid migration and to understand local patterns of fluid discharge at the eastern flank of the Juan de Fuca Ridge.

4.10 Conclusions

Seismic data sets with different source frequencies were combined to present an integrated interpretation of pronounced reflectivity anomalies above a buried basement ridge. Lateral changes of reflection amplitude can be explained by lateral porosity variations in the sediment column. Narrow zones of higher porosity are further associated with higher permeability and provide potential pathways for fluid flow.

Since independent evidence suggests that fluid upflow above the buried basement ridge is correlated with high porosity values, and this study finds that seismic reflection amplitudes decrease with increasing porosity, it is inferred that fluid discharge in the vicinity of ODP Sites 1030 and 1031 can be imaged by seismic data.

4.11 Acknowledgments

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Chapter 5: Correlation between very high resolution seismic data from the eastern flank of the Juan de Fuca Ridge and ODP Leg 168 GRAPE density measurements

Lars Zühlsdorff and Volkhard Spieß, to be submitted to Journal of Geophysical Research.

5.1 Abstract

At the eastern flank of the Juan de Fuca Ridge, single channel and multi-channel very high resolution seismic data were recorded with a water gun and two GI-Guns, which provided different source frequencies. To ground truth the seismic records, traces were correlated with synthetic seismograms calculated from GRAPE density measurements on ODP Leg 168 drill cores. The quality of both core logging and seismic data is sufficiently high to reveal a relation between density variations and seismic reflections at source frequencies above 100 Hz. This suggests that density approximates subbottom variations in seismic impedance. GRAPE density is also sufficient to model a frequency-dependent decrease of seismic reflection amplitudes above a buried basement ridge, suggesting that low reflectivity is associated with lateral in-situ variations in sediment density and porosity, which appear to be related to fluid upflow. The correlation between seismic records and core data further reveals distinct depositional environments, which are divided by the buried basement ridge. Only one seismic reflector can be traced throughout all seismic lines, indicating that accumulation rates for the youngest part of the sediments east and west of the ridge are not very different. In the western part of the study area, a transition from an older to a younger part of the sediment section is observed. The former is characterized by focused turbiditic input in the vicinity of ODP Site 1024 and a gradual decrease in grain size towards ODP Site 1025. In the younger part, however, much thinner turbidites were deposited more evenly between both sites.

5.2 Introduction

Extraction of geologic information from a seismic record usually requires correlating seismic data with well data. This can be accomplished by a visual correlation of seismic traces with synthetic seismograms, which are generated from velocity and density data

of a nearby drill site. In the past, synthetic seismograms were successfully used to ground truth seismic reflection data, to provide insight into the large-scale velocity and density structure of sedimentary sequences, and to determine the nature of individual reflectors [e.g., *Shipley*, 1983; *Mayer et al.*, 1986; *Mosher et al.*, 1993; *Rohr and Gröschel-Becker*, 1994]. The best results are commonly achieved when downhole logging data from near-vertical wells are available [e.g., *White and Hu*, 1998]. If no downhole velocity log exists, it is possible to estimate sonic data from a density log [e.g., *Adcock*, 1993]. However, if downhole logging data are incomplete, of bad quality, or just missing, velocity and density profiles can also be constructed from measurements on rock samples or from core logging data as provided by a P-wave logger (PWL) and the Gamma Ray Attenuation Porosity Evaluator (GRAPE).

However, synthetic seismograms calculated from well data often do not agree well with seismic data [e.g., Simmons and Backus, 1996]. This can be attributed to uncertainties associated with the acquisition and processing of both seismic data and sediment physical properties. In the absence of downhole logs, the depth scale of measured core data is uncertain due to incomplete recovery or core compression or expansion during retrieval and storage. Furthermore, measured density and velocity values may be erroneous (e.g. near the ends of core sections) or biased with respect to an oversampling of specific lithologies (e.g. mud vs. sand). Seismic data on the other hand may be deteriorated by structural complexity, by steep lateral gradients in velocity or reflectivity, or by finely layered sequences causing interference due to impedance fluctuations [White and Hu, 1998]. The comparison of well data and seismic records is further complicated by the fact that the core data are measured as a function of depth whereas seismic data are recorded as a function of two-way travel time (TWT). Conversion between depth and time requires exact knowledge of the in-situ velocity profile, which is rarely available [Mayer et al., 1986]. Also, physical properties measurements are taken at discrete locations representing only a small sediment volume, whereas individual seismic amplitudes integrate over a larger sediment volume by a superposition of reflections. Thus, in many cases, obtaining an accurate well tie may not be easy. However, if both well data and seismic records are of good quality, it should be possible to successfully model the seismic response and fit both data sets in the vicinity of a drill site [e.g., Mayer et al., 1985; White and Hu, 1998]. At the very

least, a correlation between seismic and well data should be satisfactory in most marine environments and can even be very good in some areas [White and Hu, 1998].

The eastern flank of the Juan de Fuca (JdF) Ridge is a suitable location for a combined interpretation of well data and seismic records since massive turbiditic sediment input from the continent provides sharp impedance contrasts and strong reflectors, which can be traced throughout large areas. Multi-sensor track (MST) core logging provides closely spaced density and velocity data from Ocean Drilling Program (ODP) Leg 168 drill sites [*Shipboard Scientific Party*, 1997b, c]. Furthermore, the westernmost six ODP sites are connected by a dense grid of very high-resolution seismic data with different source frequency ranges, which were presented by *Zühlsdorff et al.* [1999] (Chapter 4). In this study, the three main objectives are (1) to validate that seismic reflections are associated with subsurface changes of physical properties and can thus be related to geology, (2) to investigate the nature of a frequency dependent decrease in reflection amplitude above a buried basement ridge, and (3) to derive information about time dependent changes of sediment accumulation east and west of the buried basement ridge.

5.3 Study area

ODP Leg 168 was targeted to elucidate the fundamental physics and fluid chemistry of ridge-flank hydrothermal circulation and the consequent alteration of the upper igneous crust and sediments that host the flow [*Shipboard Scientific Party*, 1997a]. The six westernmost sites are situated in the vicinity of an 1.4 Ma old buried basement ridge about 40 km east of the JDF ridge crest (Figure 5.1). Fluid discharge was inferred from geochemical pore-fluid analyses at sites of minimum sediment thickness above elevated basement [*Wheat and Mottl*, 1994; *Shipboard Scientific Party*, 1997b].

The relief of the ridge confined a sediment distributary system early in the depositional history of the area [*Shipboard Scientific Party*, 1997b]. The sediments consist of sand-rich and silt-rich Pleistocene turbidites up to 1 m thick interbedded with hemipelagic mud typically tens of centimeters thick [*Shipboard Scientific Party*, 1997a]. In the western part of the study area, seafloor bathymetry suggests that the turbidites enter the area from the north via four distributary channels (Figure 5.1). Two



Figure 5.1 Location of Ocean Drilling Program (ODP) Leg 168 sites and multichannel seismic lines presented in this paper (bold segments). Supplementary single channel data are presented for Lines GeoB96-094 and GeoB96-104. Seafloor bathymetry was provided by the Hydrosweep multibeam swath sounder (contour levels 5 m), indicating the position of possible sediment transport channels.

of the channels merge near ODP Site 1024 and meet the other two further to the south [Shipboard Scientific Party, 1997c].

At ODP Sites 1030 and 1031 above the basement ridge, sediments are characterized by a relatively high carbonate content and very little sand [*Shipboard Scientific Party*, 1997b]. Porosity profiles lack the compaction gradients observed near the seafloor at all adjacent ODP sites. It could not be determined from core studies whether the observed porosity profiles reflect active pore-water upwelling or differences in lithology [*Shipboard Scientific Party*, 1997b].

In 1995, single channel seismic data were collected onboard R/V *Tully* in preparation of ODP Leg 168. In addition, a high-resolution multi-channel seismic survey was carried out during R/V *Sonne* Cruise SO111 [*Villinger et al.*, 1996], shortly after ODP Leg 168 was finished. The objective was to identify and to investigate fluid upflow zones as predicted by *Wheat and Mottl* [1994] and to image associated sedimentary structures. The main results were presented by *Zühlsdorff et al.* [1999] (Chapter 4). In this study, the multi-frequency seismic data are compared to synthetic seismograms calculated from ODP Leg 168 core logging measurements.

5.4 Acquisition and processing of seismic and hydroacoustic data

5.4.1 Single channel seismic data

Single channel data had been collected during a R/V *Tully* cruise in 1995 [*Davis et al.*, 1997b]. A Generator-Injector gun source (GI-Gun #1, 2 x 1.5 L) was used, which provided a frequency range of about 20-120 Hz. A shot rate of 10 s corresponded to a shot spacing of 25 m. All 16 hydrophone groups of a 100 m long Teledyne streamer were stacked and traces were recorded at a sampling rate of 2 ms. A brief description of standard processing is given below and in Table 5.1.

5.4.2 Multi-channel seismic data

The GeoB multi-channel seismic system of the University of Bremen was designed to optimize lateral and vertical resolution. The data acquisition unit was specially modified to work at high shot and sampling rates. In this survey, shots were recorded every 10 s at a sampling rate of 125 μ s. A 300 m long Syntron streamer, equipped with separately programmable hydrophone subgroups, was optimized for the water depth of > 2400 m using 24 channels at a hydrophone group length of 6.25 m. Four remotely controlled

birds kept the streamer depth within a range of 1 m, and magnetic compass readings allowed to determine the position of each streamer group relative to the ship's course. Two seismic sources, a GI-Gun (GI-Gun #2, $2 \times 0.4 \text{ L}$) and a Sodera S 15 water gun (0.16 L), were operated in an alternating mode and provided frequency ranges of 100-500 Hz and 200-1600 Hz, respectively.

Properties of the seismic data sets and basic processing steps are summarized in Table 5.1. For multi-channel data, custom software was used to correct for streamer towing depth and to determine geographic positions of common depth points (CDPs) at a spacing of 20 m, accounting for streamer drift and variations in ship velocity.

	GI-Gun #1	GI-Gun #2	water gun	Parasound
Bandwidth, kHz	0.02-0.12	0.1-0.5	0.2-1.6	2.0-6.0
Radius of Fresnel zone*, m	176 (at 62.5 Hz)	88 (at 250 Hz)	44 (at 1000 Hz)	22 (at 4000 Hz)
Trace spacing, m	25 (trace spacing)	20 (bin spacing)	20 (bin spacing)	5 (average trace spacing)
Fold	1	9	9	1
Processing	filter, migration	filter, CDP-stack, migration	filter, CDP-stack, migration	filter

Table 5.1 Properties of data sets

*Calculated for a depth of 2600 m.

Both single channel and multi-channel data were carefully processed to preserve all amplitude information. Only spherical divergence was corrected by multiplying with travel time and assuming a constant velocity for each trace. Since the streamer is short compared to the water depth, an error of about 10% was expected for estimates of seismic velocities. However, velocity analysis revealed only small deviations from a value of 1500 m/s. A constant velocity model is in good agreement with velocity profiles from the ODP sites [*Shipboard Scientific Party*, 1997b, c] and was initially used for time migration of seismic data. As described later in this study, a correlation of seismic data with synthetic seismograms resulted in a slight modification of the initial constant velocity model below 40 m sub-bottom depth, but reprocessing of the data confirmed that the modification has no significant effect on the migrated seismic sections.

5.4.3 Hydroacoustic data

Two hydroacoustic systems were used simultaneously along the multi-channel seismic lines. The narrow-beam Parasound echosounder with an opening angle of 4° operated at 4 kHz and provided detailed acoustic images of the upper 50-100 m of the sediments.

The Hydrosweep multibeam swath-sounder was used to map the seafloor bathymetry as displayed in Figure 5.1.

5.5 Synthetic seismograms

In the convolutional model, a seismic trace can be modeled by convolving a seismic source wavelet with the subsurface reflectivity. The wavelet then represents a filter, and consequently the synthetic seismogram can be considered as a filtered, noise-free version of the reflectivity function. However, since in most cases the source wavelet can not be determined, it must either be extracted from seismic data or a synthetic wavelet of defined shape and bandwidth has to be used. The reflectivity function on the other hand is approximated by a discrete time series of reflection coefficients, which are derived from physical properties core logging measurements. For synthetic seismograms presented in this study, normal incidence plane waves are considered and attenuation effects are neglected.

5.5.1 Reflection coefficients

The normal incidence reflection coefficient for plane waves and between media of densities ρ_1 and ρ_2 and velocities v_1 and v_2 is given by $R=(v_2\rho_2-v_1\rho_1)/(v_2\rho_2+v_1\rho_1)$. In the absence of downhole logs, closely spaced (5 cm) GRAPE density and PWL velocity measurements from ODP Leg 168 core logging can be used for the calculation of acoustic impedance (i.e. the product of density and velocity) as a function of depth.

GRAPE density data are available for all ODP sites within the study area and are systematically higher than density values measured on discrete index property samples [*Shipboard Scientific Party*, 1997c]. This can be attributed to the onboard processing software used during ODP Leg 168. A correction of the offset would require detailed knowledge of the grain densities of the material measured [*Shipboard Scientific Party*, 1997c] and was not applied since relative density changes will not be significantly affected. However, a moving window of 1 m length was used to detect and to eliminate erroneous density values, which are significantly lower than the local average and can be assumed to be related to voids, cracks, or gas bubbles within the cores and near the ends of core sections. In addition, data gaps were linearly interpolated to avoid unrealistically high reflection coefficients across gradually changing impedance variations. Both corrections were routinely applied to all GRAPE data (Figure 5.2), but



Figure 5.2 GRAPE density core logging data used to calculate seismic reflection coefficients. A moving window of 1 m length was used to detect and to eliminate erroneous values, which are significantly lower than the local average and can be assumed to be related to voids, cracks, or gas bubbles within the cores and near the ends of core sections. In addition, data gaps were linearly interpolated to avoid unrealistically high reflection coefficients across gradually changing impedance variations. Both corrections were routinely applied but there was no further modification or tuning of individual values.

there was no further modification or tuning of individual values. A correction to in-situ conditions, which accounts for temperature change, decrease in hydrostatic pressure, and porosity rebound during core recovery [*Hamilton*, 1965, 1976] was not carried out since relative density variations within the upper 150 m of the sediment column will not be significantly changed [e.g., *Mosher et al.*, 1993; *Rohr and Gröschel-Becker*, 1994].

PWL velocity data of ODP Leg 168 are less important and useful for the calculation of reflection coefficients since values strongly scatter around a mean trend. Deviations from the average are higher than typical relative velocity variations in marine surface sediments, which do not exceed 5%, especially if porosity is > 50% [e.g., *Weber et al.*, 1997; *Breitzke*, 1999]. In contrast, the associated changes in density within terrigeneous sediments are usually > 20% [e.g., *Weber et al.*, 1997; *Breitzke*, 1999]. Therefore, seismic reflection coefficients are rather dominated by density than by velocity and can approximately be calculated for variations in GRAPE density only. The error, which is introduced by assuming a constant velocity and ignoring the velocity contribution to variations in reflection coefficients, is estimated to be only 10-20%. However, as described below, mean velocity trends can still be used to convert the depth scale of the density values into a time scale, which is important to correlate variations of GRAPE data with seismic reflectors.

The impulse response function for seismic modeling was computed using the state space approach [*Mendel et al.*, 1979], which takes all possible internal multiple reflections into account. For comparison, reflection coefficients were also calculated directly from the GRAPE density log. The results are almost identical at all sites, suggesting that the influence of multiple reflections is negligible for the data presented in this study.

5.5.2 Wavelets

To produce synthetic seismograms, two different types of wavelets were used, namely synthetic wavelets as described by *Ricker* [1953] as well as wavelets which were extracted from recorded seismic traces. Alternatively, application of a standard bandpass filter to the reflectivity function also produces a synthetic seismogram.

Assuming that the impulse response of the earth can be considered as "white noise", the amplitude spectrum of a seismic trace is dominated by the amplitude spectrum of the seismic source wavelet. Thus, zero-phase wavelets as shown in Figure 5.3a-d can be

estimated from amplitude spectra of GI-Gun, water gun, and Parasound traces. Before they were plotted, all wavelets were first normalized with respect to their maximum amplitude. For comparison, Ricker-wavelets calculated for center frequencies of 60, 200, 400, and 4000 Hz, respectively, are shown in Figure 5.3e to illustrate the differences in signal length and temporal resolution.





The wavelets in Figures 5.3a and 5.3b are very similar to seafloor reflections, which were picked from GI-Gun data. In contrast, picking of typical first arrivals from water gun and Parasound data is more difficult, which is mainly due to their lateral variability. Seafloor reflections were therefore not used for the calculation of synthetic seismograms presented in this study.

5.5.3 Correlation to seismic records

Before synthetic seismograms were generated, the depth scale of the reflection coefficients had to be converted into two-way travel time. Initially, this conversion was carried out using a constant velocity of 1500 m/s for all drill sites. A sampling rate of only 25 μ s was chosen to preserve the complete information provided by GRAPE data. The reflection coefficients were then convolved with different wavelets, resampled, and compared to seismic traces in the vicinity of the corresponding ODP sites. It was assumed that a correlation between seismic records and GRAPE density variations can be best achieved when the velocity profile is precisely matching and consequently the time scale of the reflection coefficients is correct. Thus, general misfits increasing with

depth can be attributed to deviations from correct interval velocities and were minimized by modifying the original constant velocity profile. However, velocities were only modified within a range of realistic values according to mean trends and variations in PWL data. As a result, only misfits < 1-2 ms could be corrected. All remaining individual offsets were assumed to be associated with inaccuracies in recording geometry and processing.

5.6 Observations

5.6.1 Velocity profiles

The final velocity profiles at ODP Holes 1023A, 1024B, 1025B, 1030B, 1031A, and 1028A, which were used to convert the depth scale of reflection coefficients into travel time, are shown in Figure 5.4. Within the upper part of the sediment section, interval velocity appears to be constant at most drill sites. At sub-bottom depths > 40 m, however, values increase up to 1650 m/s as found at ODP Site 1024. At the same site, turbidites between 40 and 100 m sub-bottom depth are thicker (up to 128 cm) than elsewhere in the study area [*Shipboard Scientific Party*, 1997c]. At all other sites, interval velocity and turbidite thickness are also positively correlated.

5.6.2 Effects of wavelet type and frequency content

In Figure 5.5, a detailed comparison of different synthetic seismograms is presented, which were calculated for ODP Site 1024. Different types of wavelets were used, i.e. a Ricker-wavelet of 200 Hz center frequency (Figure 5.5a), the wavelet shown in Figure 5.3b (Figure 5.5b), and a standard bandpass filter (100-500 Hz) with a moderate taper (Figure 5.5c). In Figure 5.5d, the synthetic seismograms from Figure 5.5a are further compared to recorded traces of Line GeoB96-084 (GI-Gun #2). Independent of the type of wavelet used, almost all reflectors in the vicinity of ODP Site 1024 can be correlated in detail to variations in GRAPE density, indicating that density approximates subbottom changes in seismic impedance.

For the same site, Figure 5.6 illustrates that correlation of model traces with water gun data is more complicated. One reason is that seismic attenuation effects are more visible at higher source frequencies. Thus, synthetic seismograms of lower frequency content (Figure 5.6a, Ricker-wavelet, 400 Hz center frequency) are more suitable to be compared to deep reflections, whereas synthetic seismograms of higher frequency



Figure 5.4 The final velocity profiles used to convert the depth scale of reflection coefficients into travel time. An increase in interval velocity is observed below 40 m sub-bottom depth, which is in agreement with mean trends in P-wave logger (PWL) measurements and increasing turbidite thickness.



Figure 5.5 Synthetic seismograms calculated from GRAPE density data of ODP Site 1024 using (a) a Ricker-wavelet (200 Hz center frequency), (b) the wavelet shown in Figure 5.3b, and (c) a bandpass filter (100-400 Hz) with a moderate taper. (d) The traces shown in (a) are compared to multichannel seismic data (Line GeoB96-084, GI-Gun #2). The complete seismic line is presented in Figure 5.11a. Synthetic seismograms in this and in the following figures were fivefold plotted to facilitate visual correlation of individual reflections. Independent of the type of wavelet used, observed seismic reflections can be related to variations in GRAPE density, indicating that density approximates subbottom variations in seismic impedance.
content (Figure 5.6b, wavelet as shown in Figure 5.3c) better model the characteristics of seismic reflections in the upper part of the sediment section (e.g., at 3.6 s), where overall attenuation is weaker. Visual correlation is further complicated by a large number of reflectors, which may also reveal a more pronounced lateral variability with respect to reflection time and amplitude. Even at higher resolution, however, most seismic reflections can still be related to density variations within the sediments.

The synthetic seismograms presented in all following figures were produced using Ricker-wavelets as shown in Figure 5.3e. In Figure 5.7, single channel and multichannel data of Line GeoB96-094 (ODP Site 1028) are compared. The correlation between modeled and recorded traces is much better for the higher source frequency range of GI-Gun #2 (Figure 5.7b) than for the frequency range provided by GI-Gun #1 (Figure 5.7a). Only few misfits at 3.70-3.74 s are observed, which may be explained by the presence of structural complexities in the vicinity of the drill hole. Modeled water gun data (Figure 5.7c) provide the best temporal resolution and allow to detect gaps in GRAPE measurements, which cannot be identified with GI-Gun data (compare to Figure 5.2 and next Figure 5.8). Slight variations in travel time to the seafloor are due to differences in triggering and signal generation of seismic sources.

For the same reflectivity function, the effect of a change in source frequency from 20 to 1600 Hz is shown in greater detail for a Ricker-wavelet (Figure 5.8). Most of the stronger reflections in modeled traces can be identified at all frequencies above 100 Hz, although the character of reflection patterns significantly changes with increasing resolution (e.g. at 3.68 s). Below 100 Hz, however, temporal resolution decreases rapidly and travel time for strong reflections is not consistent for varying frequency. For comparison, the source frequency ranges of data presented in this paper are given in Table 5.1.

5.6.3 Reflection amplitudes

In the vicinity of ODP Site 1030, single channel, multi-channel, and echosounding data of Line GeoB96-104 are compared (Figure 5.9). Zones of low reflection amplitudes are observed, which extend from the basement ridge to the seafloor. The amplitude decrease appears to be frequency dependent and is most pronounced at higher source frequencies as provided by water gun and Parasound sources (Figures 5.9a-d). As a result, Parasound amplitudes at ODP Site 1030 are very small and a comparison with



Figure 5.6 Comparison of multichannel seismic data (Line GeoB96-084, water gun) and synthetic traces calculated for ODP Site 1024 using (a) a Ricker-wavelet (400 Hz center frequency), and (b) the wavelet shown in Figure 5.3c. The complete seismic line is presented in Figure 5.11b. The characteristics of observed seismic reflections in the upper part of the sediment section (e.g., at 3.6 s) are better modeled by using the wavelet shown in Figure 5.3c, whereas the Ricker-wavelet of lower frequency content is more suitable to model deeper reflections, which are more affected by attenuation effects.



Figure 5.7 Synthetic traces calculated for ODP Site 1028 using Ricker-wavelets with center frequencies of 60, 200, and 400 Hz, respectively, are combined with seismic records (Line GeoB96-094): (a) Single channel data (GI-Gun #1), (b) multichannel data (GI-Gun #2, compare to Figure 5.12a), and (c) multichannel data (water gun, compare to Figure 5.12b). The least discrepancies between modeled and recorded traces are observed at the frequency range provided by GI-Gun #2 (Table 5.1).



Figure 5.8 Synthetic traces calculated for ODP Site 1028 using Ricker-wavelets of various center frequencies. Two gaps in GRAPE measurements exist at about 3.615 and 3.755 s TWT, respectively. Most of the stronger reflections can be identified for frequencies above 100 Hz, although the character of reflection patterns significantly changes with increasing resolution (e.g. at 3.68 s). Below 100 Hz, time resolution decreases rapidly and travel time for strong reflectors is not consistent for varying frequency. See Table 5.1 for the source frequency ranges of data presented in this study.



Figure 5.9 Single channel and multichannel seismic data (Line GeoB96-104) across the buried basement ridge and ODP Site 1030: (a) GI-Gun #1, (b) GI-Gun #2, (c) water gun, and (d) Parasound. Vertical exaggeration (VE) for Figures 5.9, 5.11, and 5.12 was calculated assuming a constant velocity of 1500 m/s. Reflector A can be traced throughout the study area (compare to Figures 5.11 and 5.12). Above the basement elevation, a decrease of reflection amplitudes is observed which is frequency dependent and most pronounced at higher source frequencies. Thus, Parasound amplitudes at ODP Site 1030 are very small and a comparison with synthetic traces was not attempted.

synthetic traces was not attempted.

Frequency dependent amplitude anomalies are observed in all multi-channel seismic lines across the buried basement ridge [Zühlsdorff et al., 1999]. Cores from ODP Sites 1030 and 1031 above the ridge contain a relatively small proportion of sand and do not reveal a compaction-related decrease of porosity with depth. At these sites, also the lowest bulk densities compared to all other ODP Sites within the study area were observed [Shipboard Scientific Party, 1997b].

The effect of GRAPE density on amplitudes of synthetic seismograms is further investigated by a comparison between all ODP sites (Figure 5.10). For wavelets estimated from single channel, multi-channel, and Parasound records (Figures 5.3a-d), root-mean-square (rms) amplitudes were calculated for the upper 50 ms below sea floor and normalized to the average value of each source type. With one exception in Figure 5.10a, the lowest rms-amplitudes are observed at ODP Sites 1030 and 1031, suggesting that low reflection amplitudes are associated with a local reduction in density contrasts. Furthermore, the effect appears to be dependent on signal length since the strongest amplitude decrease is observed for the wavelets with the highest frequency content, whereas the variation of rms-amplitudes in Figure 5.10a (GI-Gun #1) is relatively small.



Figure 5.10 Variation of root-mean-square (rms) amplitudes of synthetic seismograms, which were generated using (a) GI-Gun #1, (b) GI-Gun #2, (c) water gun, and (d) Parasound wavelets as shown in Figure 5.3a-d. Rms-amplitudes were calculated for a time window of 50 ms and normalized to the average value of each data set. With one exception in Figure 5.10a, the lowest rms-amplitudes are observed at ODP Sites 1030 and 1031, suggesting that the zones of low reflection amplitudes (Figure 5.9) are associated with a local reduction of density contrasts.

5.6.4 Seismic stratigraphy

In addition to Line GeoB96-104 across the buried basement ridge (Figure 5.9), Lines GeoB96-084 west of the ridge and GeoB96-094 east of the ridge are presented in Figures 5.11 and 5.12, respectively. The sediment section down to the basement was subdivided into different lithostratigraphic units and subunits, i.e. Subunit IA, which includes discrete sand beds, Subunit IB, which contains silt turbidites, and Unit II, which only consists of hemipelagic mud deposits [Shipboard Scientific Party, 1997b, c]. The boundary between Subunits IA and IB is characterized by a decrease in reflection amplitudes in water gun data (Figures 5.11b and 5.12b). In sediment cores, the base of Subunit IA occurs at 121 m (ODP Site 1024), 70 m (ODP Site 1025), and 81 m subbottom depth (ODP Site 1028), which is in agreement with seismic reflector depths for the velocity profiles shown in Figure 5.4. The nature of the boundary between the subunits can be examined in Figure 5.11, where seismic reflectors assigned to Subunit IA at ODP Site 1024 can be traced to a sub-bottom depth at ODP Site 1025 which is part of Subunit IB. Turbidite thickness at ODP Site 1024 reaches a maximum of 128 cm, but is smaller (30-50 cm) at ODP Site 1025 [Shipboard Scientific Party, 1997c]. Thus, the boundary between the subunits in the western part of the study area appears to be associated with lateral variations in average turbidite properties and can not be assigned to a single seismic reflector.

A link of seismic stratigraphy between the western and eastern parts of the study area has not been established since seismic reflectors can not be traced with confidence across the buried basement ridge due to lower sedimentation rates, inclined bedding, and low reflection amplitudes. The only connection is provided by a seismic line south of the main ridge where topography is moderate (Figure 5.1). However, seismic reflectors, which are strong in the vicinity of ODP Site 1028, completely disappear along the line towards the western ODP sites. Vice versa, reflectors identified in the western part of the study area can not be traced far into the eastern part. Only one seismic reflector, which in this study is called "Reflector A", can be traced throughout the study area (Figures 5.9b, 5.11a, and 5.12a). At the location of the ODP sites, Reflector A is observed in sub-bottom depths of approximately 21 m (ODP Sites 1023, 1024, and 1025), 15 m (ODP Site 1030), and 25 m (ODP Site 1028) with an error of 1-2 m. At ODP Site 1031, the depth of Reflector A is estimated to be 12 m, but the error



Figure 5.11 Multichannel seismic data (Line GeoB96-084) across ODP Sites 1024 and 1025: (a) GI-Gun #2, and (b) water gun. A 10 m thick gap in GRAPE measurements exists at about 3.64 s TWT (ODP Site 1025). Reflector A can be traced throughout the study area (compare to Figures 5.9 and 5.12). The boundary between Subunits IA and IB is characterized by a decrease of water gun amplitudes and can not be associated with a single seismic reflector.



Figure 5.12 Multichannel seismic data (Line GeoB96-094) across ODP Site 1028: (a) GI-Gun #2, and (b) water gun. The boundary between stratigraphic Subunits IA and IB can be identified by a decrease of water gun amplitudes. Reflector A can be traced throughout the study area and provides a link between the seismic sections east and west of the buried basement ridge (compare to Figures 5.9 and 5.11). may be larger since seafloor topography is rough, reflectors are steep and converging and reflection amplitudes are lower, and thus, an accurate correlation of modeled traces with seismic records is more difficult at this site.

An assignment of ages to seismic reflectors is only possible in the western part of the study area. Calcareous nannofossil events at 0.09 and 0.28 Ma, which were determined on cores from ODP Sites 1023 and 1024 [*Shipboard Scientific Party*, 1997c], can be linked to seismic reflectors. However, the depth of those reflectors is not consistent with the depth of the corresponding nannofossil events at ODP Sites 1025 and 1030. Assuming that the age profiles at ODP Sites 1023 and 1024 are correct, Reflector A is much younger than 0.09 Ma.

5.7 Discussion

As stated above, the main objectives of this study are to ground truth high resolution seismic data, to examine anomalies in reflection amplitude above the basement ridge, and to reveal information about spatial and temporal variations of sediment accumulation. With respect to these subjects, the observations described in the previous section are discussed below.

5.7.1 Ground truthing of seismic data

Figure 5.8 illustrates that the success of correlating synthetic seismograms and recorded seismic traces depends on the source frequency range. At dominant frequencies below 100 Hz, only few seismic reflections are reproduced from the series of reflection coefficients and their corresponding reflection time does not appear to be reliable. It is assumed that most of the misfits observed in Figure 5.7a (GI-Gun #1) are due to the low main frequency of about 60 Hz.

For increasing dominant frequencies above 100 Hz, the number of details, which can be resolved, increases rapidly, although small misfits may complicate a visual correlation between modeled and recorded traces. Inaccuracies of recording geometry and processing as well as small deviations in the velocity model and the stretching of the source signal with travel time due to seismic attenuation are more critical within the frequency range provided by the water gun than for lower source frequencies (Figures 5.6 and 5.7c). However, most reflections in water gun data can clearly be related to subsurface variations in density. The frequency range provided by GI-Gun #2 reveals the least discrepancies between recorded and modeled traces within the upper 200 m of the sediment section (Figures 5.5d and 5.7b). Almost all seismic reflections can be modeled from variations in density, implying that both seismic and GRAPE data are of good quality, processing of density values was successful, and density dominates downhole acoustic impedance variations.

As a result, the multi-channel seismic data presented in this study can be directly related to borehole data and can therefore be interpreted with respect to lithology.

5.7.2 Amplitude anomalies above the basement ridge

Since bulk density of saturated sediments is directly related to porosity, and porosity thereby influences seismic impedance, lateral porosity changes as observed above the basement ridge may be the main reason for anomalies in recorded seismic reflection amplitudes [Zühlsdorff et al., 1999]. Thus, variations in sediment porosity and density between different ODP Sites should also be expressed in reflection amplitudes of synthetic seismograms.

However, a comparison of modeled and recorded reflection amplitudes is not straightforward since the contribution of velocity to seismic impedance had to be neglected during modeling and depth dependent attenuation effects were ignored. Furthermore, density values only represent the sediments within the drill hole whereas the amplitude value at each recorded time sample is associated with a larger sediment volume. Therefore, the interpretation of modeled reflection amplitudes is limited to a comparison of relative changes of rms-amplitudes between different ODP Sites. Locally reduced impedance contrasts, and thus low rms-amplitudes, are expected at ODP sites which reveal higher average downhole porosities [*Zühlsdorff et al.*, 1999].

In fact, the lowest rms-amplitudes for each seismic source are observed at ODP Sites 1030 and 1031 (Figure 5.10), which reveal higher porosities and lower densities than all other sites at comparable depth [*Shipboard Scientific Party*, 1997b, c]. Furthermore, the decrease in rms-amplitudes is more pronounced at higher source frequencies. For deeper reflections in recorded seismic data, this may partly be explained by attenuation effects, which are more effective at higher frequency and higher porosity [*Zühlsdorff et al.*, 1999]. However, synthetic seismograms suggest that, independent of attenuation effects, the amplitude decrease depends on the source signal length.

As a result, both seismic records and synthetic seismograms reveal locally reduced reflection amplitudes in the vicinity of the basement ridge (Figures 5.9 and 5.10). Thus, the frequency-dependent amplitude decrease reflects lateral changes of sediment porosity and density and cannot be assigned to processing artifacts or rough topography. Since independent evidence suggests that fluid upflow above the basement ridge is correlated with high porosity values [*Wheat and Mottl*, 1994; *Shipboard Scientific Party*, 1997b], it was inferred that fluid discharge in the vicinity of ODP Sites 1030 and 1031 can be imaged by seismic data [*Zühlsdorff et al.*, 1999].

5.7.3 Temporal and spatial variations of sediment accumulation

Tracing of seismic reflectors across the buried basement ridge is complicated, even at locations where topography is moderate. This supports the assumption that the relief of the ridge confined a sediment distributary system early in the depositional history of the area [*Shipboard Scientific Party*, 1997b]. Only one seismic reflector (Reflector A) can be traced throughout the study area (Figures 5.9, 5.11, and 5.12). However, the reflection patterns above Reflector A are different at both flanks of the ridge (Figure 5.9b), suggesting that the sedimentation regimes east and west of the ridge are different and strong events affecting the whole area are exceptional.

The sub-bottom depth of seismic reflectors at the location of the drill sites can be determined from reflection time using the velocity profiles presented in Figure 5.4. With an estimated error of < 2 m, the depth of Reflector A is 21 m at the western ODP sites and 25 m at ODP Site 1028. This indicates that accumulation rates for the youngest part of the sediment section are not much different east and west of the ridge. Above the ridge, accumulation rates are lower since the proportion of suspended material in a turbiditic particle cloud, which can reach elevated areas, predominantly consists of finer grains.

The thickest turbidites within the study area were observed at ODP Site 1024 [Shipboard Scientific Party, 1997c]. However, turbidite thickness appears to decrease eastward along Line GeoB96-084 (Figure 5.11) and is much smaller at ODP Site 1025 at comparable depth. This is in agreement with the interval velocity function, which is positively correlated with turbidite thickness, indicating that the main factor controlling seismic velocity is grain size. However, higher interval velocities are only observed below a sub-bottom depth of 40 m (Figure 5.4). This implies that one or both of the

potential turbidite channels, which merge near ODP Site 1024 (Figure 5.1), once were more active or provided coarser material than they do today. Furthermore, focused turbidite input near ODP Site 1024 and a gradual decrease of grain size in an eastward direction is consistent with the observation that the boundary between lithostratigraphic Subunits IA and IB is not associated with a single turbiditic event, indicating that the first sandy turbidite was accumulated much earlier at ODP Site 1024 than at ODP Site 1025.

As a result, a combined interpretation of seismic records and core logging data reveals differences in sediment accumulation for the eastern and the western parts of the study area as well as for the younger and the older parts of the sediment section between ODP Sites 1024 and 1025.

5.8 Conclusions

1. Although downhole logs are not available for the western sites of ODP Leg 168, and a correlation between seismic and core logging data is often complicated, seismic reflections can be related to variations in GRAPE density. This suggests that density rather than seismic velocity dominates sub-bottom variations in seismic impedance. The least discrepancies between seismic records and synthetic seismograms are observed at source frequencies above 100 Hz, indicating that both, the high resolution multi-channel seismic data presented in this study and core logging data provided by the ODP Leg 168 Shipboard Scientific Party, are of very high quality.

2. Narrow zones of low reflection amplitudes are observed in seismic lines across a buried basement ridge. The amplitude decrease is frequency dependent and most pronounced at higher source frequency ranges. This effect can be modeled on the basis of GRAPE density data, which are directly related to sediment porosity. Low reflection amplitudes thus reflect lateral changes of in-situ physical properties and do not have to be assigned to processing artifacts or rough topography.

3. A combined interpretation of seismic records and borehole data suggests that the buried basement ridge divides the study area into two parts which belong to different sedimentation regimes, although overall accumulation rates are not much different for the youngest sediments. The western part of the study area can further be divided into a younger and an older part. In the older part, sediment input was focused to the vicinity of ODP Site 1024 and grain size decreases gradually to the east, whereas in the younger part, much thinner turbidites were accumulated more evenly between ODP Sites 1024 and 1025.

5.9 Acknowledgements

Standard seismic processing was carried out with the Seismic Unix software package [Stockwell, 1997]. Maps were created using Generic Mapping Tools (GMT) [Wessel and Smith, 1991]. R/V Sonne Cruise SO111 and subsequent research on seismic data was funded by the German Ministry of Science and Technology (BMBF, Grant No. 03G111A/B). Support for this work was provided through the Deutsche Forschungsgemeinschaft (DFG). We thank C. Hübscher and W. Böke for their valuable contributions to the success of Cruise SO111 as well as the captain and the crew of R/V Sonne for excellent technical support. Co-Chiefs and Shipboard Scientific Party of ODP Leg 168 are gratefully acknowledged for providing high quality physical properties data and supplementary information.

Chapter 6: Implications for focused fluid transport at the northern Cascadia accretionary prism from a correlation between BSR occurrence and near seafloor reflectivity anomalies imaged in a multifrequency seismic data set

Lars Zühlsdorff, Volkhard Spieß, Christian Hübscher, Heinrich Villinger, Andreas Rosenberger, Geologische Rundschau, 1999, in press.

6.1 Abstract

A high resolution seismic survey was carried out at the accretionary prism on the continental slope off Vancouver Island, Canada. Two GI-Gun data sets with different source frequency ranges of 50-150 Hz and 100-500 Hz were combined with 4 kHz narrow beam echosounding data (Parasound). The data allow spatial correlation between a gas hydrate bottom simulating reflector (BSR) and distinct areas of high near seafloor reflectivity. An integrated interpretation of the multi-frequency data set provides insight into the regional distribution of tectonically induced fluid migration and gas hydrate formation in the vicinity of ODP Leg 146 Sites 889 and 890.

The BSR at the base of the gas hydrate stability field is observed within accreted and deformed sediments, but appears to be absent within bedded slope basin deposits. It is suggested that these basin deposits inhibit vertical fluid flow and prevent the formation of a BSR, whereas the hydraulic conductivity of the accreted sediments is sufficiently high to allow for pervasive gas migration. An elevation of the BSR beneath the flanks of a topographic high is interpreted as an indicator for local upflow of warm fluids along permeable pathways within outcropping accreted sediments.

Parasound data reveal discontinuous zones of high reflectivity at or directly beneath the seafloor, which may indicate local cementation of surface sediments. In combination with GI-Gun data, the occurrence of these reflective areas can be related to the location of slope sedimentary basins acting as hydraulic seals. It is proposed that the seals sometimes fail along faults extending beneath the BSR, leading to focused upflow of methane bearing fluid and the formation of carbonate pavements at the seafloor.

6.2 Introduction

Gas hydrate is an ice-like substance which is formed at high pressure and low temperature, consisting of a lattice of water molecules with cages containing guest gaseous molecules bonded by Van der Waals forces [e.g. *Sloan*, 1990; *MacKay et al.*, 1994]. The most common hydrate forming gas is methane [e.g. *Claypool and Kaplan*, 1974]. Since hydrates may store enormous quantities of methane in marine sediments, they are important as a possible clean energy source, may affect atmospheric composition and global climate change, and represent a potential submarine geologic hazard by decreasing the stability of slope sediments [*Kvenvolden*, 1988; *Kvenvolden*, 1994].

The stability conditions for solid hydrate mainly depend on temperature and pressure, and, to a lesser extent, on the composition of hydrate forming constituents and the salinity of the pore fluid [e.g. *Sloan*, 1990]. Since the subbottom depth to the base of the stability field is more strongly a function of temperature than of pressure, the base of solid hydrates approximately marks an isotherm. The associated seismic signature is a bottom-simulating reflector (BSR), because isotherms generally follow the seafloor, except where local changes of the temperature gradient are present [e.g. *Hyndman and Davis*, 1992; *Kvenvolden*, 1994]. The BSR can be interpreted either as the top of a low velocity zone caused by free gas beneath the layer [*Singh et al.*, 1993], or as the base of a high velocity layer of hydrate-cemented sediments above [*Hyndman and Davis*, 1992]. In many cases, however, the effect of hydrate and free gas appear to combine.

The common concentrations of organic matter in marine sediments are probably insufficient to locally generate the large volumes of methane needed for the formation of significant amounts of hydrate [e.g. *Claypool and Kaplan*, 1974; *Paull et al.*, 1994]. Alternatively, hydrates may form either through dissociation of pre-existing hydrate, if the base of the stability field moves upward [e.g. *Dillon and Paull*, 1983; *Kastner et al.*, 1995], or methane originating at greater depth is transported by upward pore fluid migration [*Hyndman and Davis*, 1992]. Even when the effects of in situ production and upward migration of methane on hydrate formation may also combine, some kind of vertical fluid transport appears to be essential [*Paull et al.*, 1994; *Gornitz and Fung*, 1994].

The occurrence of gas hydrate may therefore be diagnostic of upward fluid flow [Hyndman and Davis, 1992]. Further, local upward flow along permeable conduits was inferred from variations of BSR subbottom depth, which indicate local changes of the temperature gradient [e.g. Minshull and White, 1989; Davis et al., 1995]. More evidence for upward migration of pore fluids in accretionary prisms was provided through seismic estimates of porosity reduction [e.g. Bray and Karig, 1985; Minshull and White, 1989; Calvert and Clowes, 1991], by studies of the physical and thermal states of the sediments [e.g. Davis et al., 1990; Hyndman et al., 1993; Sample and Kopf, 1995], and by mapping and sampling of early diagenetic deposits near the seafloor [e.g. Ritger et al., 1987; Kulm and Suess, 1990; Carson et al., 1994].

Ever since dewatering processes at accretionary prisms were first investigated, some questions have remained, including the magnitude of flow, the modulation of flow rates over time and the relative importance of focused flow versus diffuse flow [Suess et al., 1998]. Upflow velocities along faults were suggested to be 2 to 3 orders of magnitude higher than dispersed flow through the sediment [Moore et al., 1990], but dispersed flow, accommodated by intergranular and small scale fracture permeability, leaks out over much larger areas and thereby causes an unknown proportion of water loss [e.g. Carson and Screaton, 1998]. An accurate estimate of mass transport rates is further hampered by the effect of transient processes, since fluid expulsion may be coupled to episodic fault displacement and to the earthquake cycle [Moore and Vrolijk, 1992]. Hence, the distribution and nature of flow must be determined at any study area before a quantitative estimate of the fluid budget becomes possible.

At the northern Cascadia accretionary prism, tectonic loading rather than tectonic compression may be the most important factor controlling excess pore pressure generation and fluid migration [Hyndman and Davis, 1992; Moran et al., 1995]. Approximately 3.5 km of Cascadia Basin sediment is almost completely scraped off the downgoing Juan de Fuca plate, and the thickness of the sediment wedge doubles over a distance of only 10-20 km from the deformation front [e.g. Davis and Hyndman, 1989; Hyndman and Davis, 1992]. One consequence of this tectonic thickening is rapid porosity loss with depth, associated with the expulsion of large volumes of fluids [e.g. Davis et al., 1990]. ODP Leg 146 was directed to the investigation of this fluid flow and sediment deformation at the Cascadia Margin [Shipboard Scientific Party, 1994].

In this study, fluid transport processes in the vicinity of ODP Leg 146 Sites 889 and 890 are investigated in greater detail. Both seafloor and subbottom features related to fluid flow are imaged in a multi-frequency seismic data set providing sufficient resolution at each depth level. Distinct areas of high seafloor reflectivity observed in Parasound data are related to seismic observations of a gas hydrate BSR. This provides insight into the regional distribution of diffuse and confined tectonically induced fluid migration and gas hydrate formation.

6.3 Study area

Ten lines of single channel and multi channel seismic data were collected within an area of about 15 by 15 km in the vicinity of ODP Sites 889 and 890 (Leg 146). The drill sites are located on the continental slope off Vancouver Island in water depths of 1315 m and 1326 m, respectively (Figure 6.1). At both locations, the pervasively fractured and folded material of the accretionary wedge is covered by a veneer of stratified slope sediments. They appear to consist predominantly of turbidites and hemipelagic deposits [*Shipboard Scientific Party*, 1994]. Figure 6.1 shows part of a 20-km-wide gently undulating plateau, where accreted sediments locally crop out. A steep escarpment to the west separates the plateau from the lower slope.

Within the study area, the absence of major thrust faults, as well as the lack of significant thermal anomalies, suggest diffuse dewatering hosted by intergranular permeability [e.g. *Hyndman and Davis*, 1992; *Kastner et al.*, 1995]. At the location of the ODP sites, however, a linear temperature decrease with depth implies conductive heat loss rather than advection, although upward fluid movement at rates of about 1 mm/yr cannot be excluded [*Shipboard Scientific Party*, 1994]. Unfortunately, the ODP packer experiment at Site 889 (Circulation Obviation Retrofit Kit, CORK) failed due to technical problems, and in situ hydrogeologic data of dispersed flow and discharge are not available [*Carson and Westbrook*, 1995].

6.4 Data acquisition and processing

The seismic survey was carried out during R/V Sonne Cruise SO111 in late summer 1996 [Villinger et al., 1996]. Single channel data along Lines 1 to 7 (Figure 6.1) and multi channel data (Lines GeoB96-069, GeoB96-071 and GeoB96-073) were collected,



Figure 6.1 Location of ODP sites, single channel seismic lines (Lines 1-7) and multi channel seismic lines (GeoB96-069, GeoB96-071, GeoB96-073). Bathymetry data were acquired with the Hydrosweep system onboard R/V Sonne (contour levels 40 m). The quadrangle indicates the working area.

providing images of the complete sediment section down below BSR depth. Simultaneously, two hydro-acoustic systems were operated along all seismic lines, imaging the upper part of the sediment column at very high resolution (Parasound) and providing a map of seafloor bathymetry (Hydrosweep). Some details about data acquisition and processing are listed below:

(1) Single channel seismics (scs): All 16 hydrophone groups of a 100 m long Teledyne streamer were stacked. A GI-Gun with 0.7 l main chamber volume and 1.7 l secondary chamber volume (GI-Gun #1) provided a frequency range of about 50-150 Hz. A shot rate of 10 s corresponded to a shot spacing of 25 m.

(2) Multichannel seismics (mcs): A 300 m long Syntron streamer, equipped with separately programmable hydrophone subgroups, was optimized for the given water depth of >1100 m using 24 groups at a length of 6.25 m. Four remotely controlled birds kept the streamer depth within a range of 1 m, and magnetic compass readings allowed to determine the position of each streamer group relative to the ship's course. A GI-Gun with 2 x 0.4 l chamber volume (GI-Gun #2) generated signal frequencies between 100 and 500 Hz.

(3) Hydro-acoustic systems: a) Depending on the type of sediment, the narrow beam Parasound echosounder (4 kHz, opening angle 4°) may penetrate up to 100 m of the sediment column. Seafloor reflections can be recorded for bedding inclinations up to 2°.
b) The Hydrosweep multibeam swath-sounder was used to map the seafloor bathymetry as displayed in Figure 6.1.

Source	Туре	Frequency, Hz	Processing
Gl-Gun #1	SCS	50-150 (bandwidth)	filter, migration
GI-Gun #2	mcs	100-500 (bandwidth)	filter, CDP-stack, migration
Parasound	echosounder	4000 (dominant freq.)	filter

Table 6.1 Properties of data sets

Standard processing for single channel and multi channel seismic data (Table 6.1) was carried out with the Seismic Unix software package [*Stockwell*, 1997]. For the multi channel data, additional custom software was developed to correct for streamer towing

depth and to determine geographic positions of CDP (common depth point) bins at a spacing of 10 m.

A seismic velocity model is needed for normal move-out correction (mcs) and time migration (scs and mcs). Therefore, velocity analysis was applied to the three multi channel lines. Due to the short streamer length, however, an error of about 10% was expected for the estimated seismic velocities. Since observed velocity variations appear to be small, a constant velocity of 1750 m/s for stacking and migration was assumed for all lines within the study area. This is in good agreement with average velocities from VSP data (ODP Site 889) [*MacKay et al.*, 1994] and from a full waveform inversion of mcs data [*Yuan et al.*, 1996]. Both data sets reveal a velocity increase from 1700 m/s to 1900 m/s in the 100 m depth interval above the BSR. This increase was attributed to a hydrate concentration of 20% of pore space just above the BSR [e.g. *Hyndman and Spence*, 1992; *Yuan et al.*, 1996]. However, since the implications and conclusions of this study are only based on relative changes of reflector strength and subbottom depth, the choice of a constant velocity model for data analysis produces negligible errors.

6.5 Observations

Within all GI-Gun data sets, the accreted and deformed material of the accretionary wedge is characterized by a total lack of coherent reflection energy. Therefore, the prism sediments can be easily distinguished from sedimentary basins, which contain stratified hemipelagic deposits (Figures 6.2a-6.5a).

A clear BSR with negative reflection polarity occurs mainly within accreted sediments at a two-way reflection time (TWT) of about 275 ms below seafloor. For the lower frequency GI-Gun #1, the amplitudes of the BSR and the seafloor reflection are of the same order (Figure 6.2a). In contrast, the BSR amplitude for the higher frequency GI-Gun #2 is just above the noise level and much smaller than the sea floor return (Figure 6.3a). For both sources, the BSR appears to be discontinuous. In Figure 6.4a, the BSR is not present between shot numbers 1 and 150 and between shot numbers 505 and 540. In both areas, basins of stratified sediments are deeper than the BSR. Figure 6.5a shows a close-up of the northern half of Line 3, which is also dominated by bedded deposits. The BSR can be identified in the vicinity of shot number 255, where the sediment section is cut by a fault, but it can not be traced throughout the basin. All seismic lines



Figure 6.2 Line 5: (a) Single channel seismic data reveal a strong bottom simulating reflector. (b) Parasound data reveal discontinuous zones of high reflection amplitudes at the seafloor. Vertical exaggeration (VE) for all figures is given for a velocity of 1750 m/s.



Figure 6.3 Line GeoB96-071: (a) Multi channel seismic data. The double reflection of the BSR is caused by the inverted wavelet of GI-Gun #2 for a negative reflection coefficient. (b) Parasound data. High seafloor reflectivity appears to be restricted to slope basin deposits.



Figure 6.4 Line 1: (a) Single channel seismic data. The BSR is missing within basins of bedded deposits but is strong within the pervasively fractured sediments of the accretionary prism. (b) Parasound data. High seafloor reflectivity appears to be restricted to slope basin deposits (compare to Figure 6.3).



Figure 6.5 Close-up of the northern half of Line 3: (a) Single channel seismic data. The BSR is observed in the vicinity of a fault. (b) Parasound data.

within the study area consistently indicate that the BSR is not observed within basins of bedded deposits, except where vertical or subvertical structures as faults are present (e.g. Figure 6.5a). In contrast, the BSR is always observed within the pervasively fractured sediments of the accretionary prism.

Parasound data reveal up to 8 m thick zones of high reflectivity at or directly beneath the seafloor, which are not imaged at the wavelengths provided by both GI-Guns (Figures 6.2b-6.5b). The zones are discontinuous and covered by a 0.5-5 m thick transparent sediment drape as shown in a close-up of the Parasound record of Line GeoB96-071 (Figure 6.6). Both, the high surface reflectivity and the kind of draping, are not typical for sediments of the Cascadia Basin [*Villinger et al.*, 1996]. Figure 6.6 also indicates that zones of high reflectivity follow the layering in the top portion, but do not reveal a sharp base. In addition, they are much thicker than the commonly observed bedded layers.





A comparison of Parasound records with GI-Gun data shows that intervals of strong surface reflectivity appear only in the vicinity of sedimentary basins (Figures 6.2-6.5). Furthermore, the occurrence of high seafloor reflectivity may be related to the occurrence of the BSR. For all lines, the locations, where a clear BSR could be identified (Figure 6.7a), and zones of high seafloor reflectivity (Figure 6.7b) were mapped. The overall pattern presented in Figure 6.7 seems to indicate that the presence of a BSR and of strong surface reflections is mutually exclusive. This relationship is most evident for the basin at the north-eastern corner of the SSR (Figures 6.4-6.5). One



Figure 6.7 Seafloor topography provided by the Hydrosweep system (contour levels 40 m), related to hand-picked locations of (a) a distinct BSR in GI-Gun data, and (b) zones of high near surface reflectivity in Parasound data. A comparison suggests that the presence of a BSR and of high reflectivity zones is mutually exclusive.

exception, where both features are present, is a small area at the northern flank of a topographic high near ODP Site 889, where sedimentary basins are shallower than the BSR (Figures 6.2-6.3).

Zones of high seafloor reflectivity, however, are not present in all sedimentary basins. GI-Gun data suggest that their occurrence may be related to the sediment structure underneath. Two types of faults, which frequently cut the bedded sediment section, can be observed: (a) growth faults (e.g. Figure 6.2a at shot number 570, Figure 6.4a at shot number 475), and (b) normal faults, which may cause several meters of vertical displacement throughout the sediment column down to >200 m subbottom depth (e.g. Figure 6.4a at shot number 25 and 445, Figure 6.5a at shot number 255). A comparison with Parasound data indicates that the occurrence of high near seafloor reflectivity may be coupled to the presence of those normal faults.

In Figure 6.8, a close-up of the deformed and faulted zone in Figure 6.2a (shot number 310), which is located at the northern flank of a topographic high, is displayed after leveling the first arrivals. The seafloor topography as well as the location of high amplitude reflectors are given in the corresponding Parasound section. A structural disturbance of the sediment column is associated with intervals of high reflectivity at the seafloor. In addition, the BSR at this location is elevated and interrupted by a gap of about 300 m width.

To determine the BSR-topography, i.e. the TWT distance to the sea floor, a map was created by leveling the seafloor reflection for all lines and then picking the two-way reflection time for the BSR. The values were gridded using algorithms of the GMT software package [*Smith and Wessel*, 1990] and plotted as gray shades of positive or negative deviation from the mean BSR depth below seafloor (Figure 6.9a). Furthermore, gray shades indicating BSR-topography were related to contour levels of the seafloor bathymetry (Figure 6.9b), suggesting that the BSR is elevated at distinct areas which correspond to the flanks of topographic highs. The maximum elevation of about 50 ms TWT (corresponding to 44 m for an assumed average velocity of 1750 m/s) is found in the small area where both a BSR and a zone of high surface reflectivity are observed, and a significant structural disturbance of the sediment column is present (Figure 6.2).



Figure 6.8 Close-up of Line 5 (compare to Figure 6.2), focusing on the BSR (lower panel) and the seafloor reflection (middle panel) after leveling the first arrivals. Parasound data (upper panel) show the location of high reflectivity zones and the original topography. The BSR is locally elevated and reveals a gap of 300 m width.



Figure 6.9 (a) Topography of the BSR. Positive values represent an elevation above an average depth below seafloor in two-way reflection time (contour levels 10 ms). White circles mark areas where a BSR is observed. (b) Topography of the BSR related to bathymetry (contour levels 40 m, compare to Figure 6.1). The maximum elevation of about 50 ms TWT is found at the northern flanks of topographic highs.

6.6 Discussion

6.6.1 Nature of the BSR

In the following discussion, the BSR is considered to mark the base of the gas hydrate stability field. This is in agreement with results from previous studies, indicating a small amount of free gas beneath a zone with 10-20% of the pore spaces filled with hydrate [*Hyndman and Spence*, 1992; *Shipboard Scientific Party*, 1994; *Kastner et al.*, 1995; *Spence et al.*, 1995; *Yuan et al.*, 1996]. Thus, the base of the hydrate layer is characterized by a relatively sharp decrease of seismic impedance and produces a BSR with reversed reflection polarity (Figures 6.2a-6.5a).

The top of the hydrate zone as well as the base of the gas layer appear to be very gradational, since no reflections can be observed even at low frequencies [Spence et al., 1995; Fink and Spence, 1999]. In addition, VSP data from ODP Site 889 indicate that the transition between hydrate cemented sediment and the gas layer below may also be gradational within a small depth interval of less than 20 m [MacKay et al., 1994]. This may explain the relative decrease of the BSR amplitude for the higher frequency signal of GI-Gun #2 (multi channel seismics). A lower BSR amplitude for higher frequency signals is consistent with results of previous studies suggesting that the BSR amplitude is frequency dependent [Spence et al., 1995; Fink and Spence, 1999]. An alternative explanation for a weak BSR in the multi channel data could be seismic attenuation. However, layers of bedded sediments still produce strong reflection amplitudes at the depth of the BSR (Figure 6.3a), implying that weak BSR amplitudes are related to interference effects rather than to an overall attenuation of reflected energy.

6.6.2 BSR distribution and sub-bottom depth

With the exception of faulted zones as shown in Figure 6.5a, no distinct BSR is observed within the slope sedimentary basins (Figures 6.2a-6.5a). Instead, the internal reflectors, which are commonly parallel or subparallel to the seafloor, reveal high reflection amplitudes. It was therefore suggested by *Spence et al.* [1995] that the BSR is still present, but may be obscured by stronger reflectors. However, the BSR is of chemical nature and is not expected to be related to the layering. Thus, it should be visible wherever other reflectors are inclined with respect to the seafloor. This is not in agreement with the data presented in this paper (e.g. Figure 6.4a, shot numbers 100-200).

Hence, the most likely explanation for the absence of the BSR in sedimentary basins is an insufficient concentration of gas and gas hydrate. Since the local production of gas in marine sediments does generally not support the formation of hydrates [e.g. *Claypool and Kaplan*, 1974], it was suggested that the occurrence of a BSR implies diffuse upward migration of gas or gas-saturated fluid [*Hyndman and Davis*, 1992]. Then, the absence of a BSR within basins of bedded deposits indicates that gas is not provided at sufficient rates.

It is therefore inferred that the bedded hemipelagic deposits inhibit significant vertical flow, whereas the pervasively fractured and permeable sediments of the accretionary prism accommodate diffuse fluid and gas migration. Consequently, the BSR should be disrupted, wherever the low permeable sedimentary basins are deeper than the base of the hydrate stability field (Figures 6.4a-6.5a). Rising gas or fluid from below may then be diverted to the sides, following the interface between the basins and the accreted sediments towards the closest permeable conduit. A local formation of hydrate in the vicinity of faults may thereby be explained by confined gas or fluid migration (e.g. Figure 6.5a, shot number 255).

The topography of the BSR was mapped in Figure 6.9, showing a distinct shallowing which corresponds to a zone beneath the northern flank of a topographic high (Figure 6.9b). In principle, this effect could be caused by a difference in seismic velocity between accreted sediments and slope basin deposits. However, no significant lateral variation of BSR depth is observed when the thickness of the bedded section laterally changes. One example is found in the southern part of Figure 6.4, where the BSR depth is not affected by the slowly increasing thickness of the bedded deposits above. This suggests that seismic velocities of accreted and bedded sediments are almost similar. Furthermore, VSP data from ODP Site 889 reveal velocities between 1700 and 1900 m/s in the accreted sediments above the BSR [*MacKay et al.*, 1994], which are not significantly higher than the basin velocity of about 1750 m/s estimated from GI-Gun data.

Alternatively, an elevation of the BSR is expected when either the thermal conductivity of the sediments or one or more of the parameters controlling hydrate stability have changed. However, thermal conductivity is assumed to be constant, since no significant change between bedded and accreted sediments was observed at ODP

Sites 889 and 890 [Shipboard Scientific Party, 1994]. Further, effects such as a local pressure decrease, local variations of salinity, or a changing composition of hydrate forming constituents may considered to be of minor importance, since hydrate stability is more sensitive to temperature variations. BSR elevations with respect to the seafloor may then be best explained by a local upflow of warm fluids, which increases the temperature gradient and rises the isotherm imaged by the BSR. Similar effects were described by *Minshull and White* [1989] for the Makran accretionary prism and modeled by *Davis et al.* [1995] for the Cascadia prism off Oregon.

On the basis of a temperature gradient of 54°C/km and a constant thermal conductivity of 1.15 W/(mK) [Shipboard Scientific Party, 1994], the depth of the BSR was used to estimate a conductive heatflow of 62 mW/m² at areas of average BSR depth. The error of this estimate may be as high as 10% due to uncertainties in seismic velocity and temperature at the BSR. However, it is in agreement with a value of 62+/-8 mW/m² given at ODP Sites 889 and 890 [Shipboard Scientific Party, 1994]. Referring to this as an average heatflow and assuming constant temperatures at the BSR, a positive heatflow anomaly of 18% may be expected in the area of maximum BSR elevation. Considering the fact that the temperature at the BSR may be lower at shallower depth and that estimation errors may be high, the heatflow anomaly may reduce to 10-15%.

As a result, a higher hydraulic conductivity favouring local thermally efficient upflow is inferred beneath the flanks of topographic highs. Although most fluid is thought to be expelled via distributed permeability, a part of the fluid loss within the study area therefore appears to be focused. However, a local variation of the regional geothermal gradient can only develop if the flow occurred recently and rapidly enough that thermal diffusion did not yet dissipate the anomaly [*Fisher and Hounslow*, 1990; *Davis et al.*, 1995]. Since it must be assumed that local rates of focused flow may well be coupled to episodes of tectonic movement, flow rates may be time-dependent and the time-average of flow is expected to be lower than it may be estimated from recent thermal anomalies [e.g. *Carson and Screaton*, 1998].

6.6.3 Near surface features

Figure 6.7b indicates that high reflectivity zones near the surface of sedimentary basins are not exclusively found at local depressions and do also occur at the flank of

the topographic high near ODP Site 889. Furthermore, they are much thicker than the commonly observed bedded layers and do not reveal a sharp base (Figure 6.6). Consequently, it is very unlikely that this kind of high surface reflectivity is related to sedimentary processes as the deposition of turbidites.

An increase in surface reflectivity may on the other hand be associated with early diagenetic processes as the cementation of surface sediment, which could increase acoustic impedance significantly. In principle, also layered gas hydrates of sufficient concentration may have a major impact on acoustic impedance. However, the BSR is not present within the sedimentary basins and distributed vertical gas migration is probably not sufficient for hydrate formation in bedded surface layers. Thus, distinct hydrate layers which extend over larger areas near the seafloor are not likely to be expected.

Another early diagenetic product previously described at the Cascadia margin are carbonate cements [e.g. Kulm and Suess, 1990; Kopf et al., 1995; Suess et al., 1998], which were attributed to the exposure of dissolved hydrocarbons to the low-temperature and low-pressure oxidizing conditions near the seafloor [e.g. Ritger et al., 1987; Carson et al., 1994]. Carbonates may precipitate when the solubility of hydrocarbon constituents in a rising fluid decreases in response to hydrostatic pressure decrease, or excess carbonate is generated by microbial methane oxidation [e.g. Kulm and Suess, 1990; Kopf et al., 1995; Suess et al., 1998]. This suggests that carbonate formation near the seafloor is often directly related to focused pore water expulsion [e.g. Ritger et al., 1987].

From the arguments given above, pore water expulsion seems to be inhibited by pronounced bedding. However, ongoing tectonic activity as the compression of the accretionary prism or major earthquakes could trigger the failure of the stratigraphic seal. Faults may then create permeable pathways between the free gas layer below the BSR depth and the surface. The rise of pore fluid and free gas is associated with an expansion of free gas, thereby increasing buoyancy forces and accelerating the upflow. Under these circumstances, gas and fluid from greater depth can rapidly mix with bottom water, resulting in chemical reactions and precipitation. Cementation of seafloor sediment layers may then cause the higher acoustic reflectivity observed in Parasound data.

Based on echosounding data, this hypothesis can not be proved in detail. It is, however, supported by seismic observations of closely spaced faults, which cut the bedded sediments (Figures 6.2a-6.5a). A fast release of pore fluids and a subsequent distribution by currents and tidal forces may further explain why zones of high reflectivity often extend over distances of about 2 km around possible vent sites (Figures 6.2b-6.5b). At the northern flank of the topographic high near ODP Site 889, Fink and Spence [1999] also described unusually high reflection coefficients at the surface within an area of about 2 km in diameter. Their analysis of synthetic seismograms revealed layer velocities and densities corresponding to those of carbonate [Fink and Spence, 1999]. Other studies documented disseminated carbonates throughout the sediment section at ODP Sites 889 and 890 [e.g. Shipboard Scientific Party, 1994; Kopf et al., 1995], although no information was available directly from the sediment surface. Hyndman et al. [1994] assumed sandy layers containing carbonates to explain the low penetration of heatflow probes around ODP Site 889. Carson et al. [1994] mapped carbonate crusts in the vicinity of incipient thrust faults seaward of the deformation front off Oregon. They also derived that the seismic impedance of the crusts is about 4-6 times higher than for hemipelagic sediments, and that they may extend for distances of more than 250 m.

Fault controlled fluid discharge through low permeable sedimentary basins is therefore assumed to explain zones of high near seafloor reflectivity. Since major thrustfaults are absent in the study area, focused upflow is mainly related to local processes. Local tectonism creates topographic depressions, which are subsequently filled with bedded hemipelagic deposits. Growth-faults are active over long periods of time and reflect the development of those sedimentary basins. However, they would not allow for continuous focused flow due to their limited vertical extent of faulting during the intermittent tectonic activity. It is more likely that episodic fault displacement along the observed normal faults, which affects the complete sediment column between the seafloor and the BSR, triggers single events of fluid expulsion, followed by periods, when the basin deposits act as an efficient seal. This is in agreement with the presence of a sediment drape, which must have been deposited after such an event (Figures 6.2b-6.5b). On the other hand, permeable connections to the seafloor must be maintained sufficiently long to allow for the creation of thick cemented zones. The thickness of cemented intervals could be potentially used to determine the duration and intensity of confined fluid release.

As a result, indications for two different types of transient focused fluid flow are present in the study area, namely episodic upflow beneath the flanks of topographic highs and fault controlled discharge at sedimentary basins. The associated time scales do not necessarily have to be identical. However, the observations of Figure 6.8 suggest, that the same tectonic event may have initiated focused upflow beneath the flank of the topographic high near ODP Site 889 as well as confined fluid release through the bedded sediments above. The elevation of the BSR with respect to the seafloor, in conjunction with the interruption by a 300 m wide gap, implies that rising warm fluids may have decomposed solid hydrates. Zones of high surface reflectivity around the vent location are in agreement with a recent study suggesting that carbonate formation at the surface and dissociation of gas hydrates may be related [*Bohrmann et al.*, 1998].

A quantitative estimate of rates of flow and mass exchange requires numerical modeling, which is beyond the scope of this paper. However, before flow can successfully be modeled, the thermal and hydraulic parameters as well as the nature and distribution of flow have to be determined. This seismic survey imaged local patterns of fluid discharge and provided a conceptual model for the fluid regime in the vicinity of ODP Sites 889 and 890, which can be used to further constrain future investigations and models.

6.7 Summary and conclusions

From previous studies and ODP Leg 146 it was known that significant volumes of gas hydrates are present at the northern Cascadia accretionary prism. The base of the hydrate zone is marked by a BSR, which can be traced within the accreted and deformed material of the sedimentary wedge. In addition, the existence of a BSR may be diagnostic of dispersed fluid flow. The fluid release of this area is in general considered to be mainly accommodated by intergranular permeability.

In this study, two GI-Gun data sets with different source frequency ranges are combined with narrow beam echosounding data and the seafloor bathymetry provided by the Hydrosweep system. An integrated interpretation provides insight into the regional distribution of tectonically induced fluid migration and gas hydrate formation in the vicinity of ODP Leg 146 Sites 889 and 890.

The thickness of the hydrate stability field can be estimated from the subbottom depth of the BSR. In distinct areas, which correspond to the flanks of topographic highs, a shallowing of the base of the hydrate zone is observed. Furthermore, the occurrence of the BSR is restricted to fractured and accreted sediments or to the vicinity of faults. In contrast, the BSR is not present within bedded hemipelagic deposits.

If active fluid transport is required to form a BSR, some characteristics of the BSR may be used to reveal information about the regional distribution of permeability. Since a clear BSR is present, diffuse fluid migration is inferred for the pervasively fractured sediments of the accretionary wedge. In contrast, low permeable bedded deposits inhibit vertical fluid flow and a BSR may only form in the vicinity of faults, where gas is locally provided at sufficient rates. The elevation of the BSR beneath the flanks of topographic highs is interpreted as a local disturbance of the thermal gradient, which is related to an increase of hydraulic conductivity within the deforming and outcropping sediments and a subsequent guided rise of warm fluids. A positive heatflow anomaly of 10-15% may be estimated for the area of maximum BSR elevation.

At the surface of bedded sediments, zones of high seafloor reflectivity are observed, which do not reveal a sharp base and are thicker than the commonly observed hemipelagic layers. Further, the occurrences of the BSR and the highly reflective zones appear to be mutually exclusive. Only few exceptions are present in areas, where basins of bedded deposits are shallower than the BSR.

Zones of high reflectivity at the seafloor are attributed to early diagenetic processes, which cement the surface sediments and increase the acoustic impedance. The precipitates probably consist of methane derived carbonates, associated with fault controlled fluid upflow through the low permeable beds of hemipelagic deposits.

In conclusion, the diffuse seepage in the vicinity of ODP Sites 889 and 890 is superimposed by at least two additional types of confined fluid release. One is guided through pathways within deforming outcrops, the second is related to faults in low permeable environments. Both types, however, are most likely related to episodic fault displacement and flow rates are considered to be time-variable. Associated time scales are not necessarily identical, although the same tectonic event may locally initiate both types of fluid expulsion. The thickness of highly reflective surface zones as well as the heatflow anomaly, which has not yet been dissipated by thermal diffusion, may be indicative for the duration and intensity of flow.

6.8 Acknowledgements

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Chapter 7: Summary and perspectives

During R/V Sonne Cruise SO111, seismic systems with source frequencies up to 4000 Hz were combined to collect high resolution multi-frequency seismic data at the Cascadia Basin and the Cascadia margin. In both study areas, the data are sufficient to reveal information about the nature and spatial extent of fluid upflow, but the flow-related features, which can be imaged within the sediments, are very different.

In the vicinity of a buried basement ridge on the eastern Juan de Fuca Ridge flank, narrow zones of low seismic reflection amplitudes are observed. The amplitude decrease is frequency-dependent and may be explained by higher porosity values. Since porosity is directly related to density, higher porosity values can locally reduce seismic impedance contrasts. In addition, Biot-Stoll's model suggests that attenuation coefficients are porosity-dependent and reflection amplitudes are further reduced at higher source frequencies when porosity increases. Since independent evidence indicates that thermally driven fluid upflow above the basement ridge is correlated with high porosity values, and this study finds that increasing porosity may decrease seismic reflection amplitudes, it is inferred that fluid upflow at this location can be imaged by seismic data.

At the northern Cascadia accretionary prism, most seismic features, which provide insight into the regional distribution of tectonically driven fluid expulsion, are related to the rise of large volumes of gas. A bottom simulating reflector (BSR) at the base of the gas hydrate stability field is observed within accreted and deformed sediments but appears to be absent within bedded slope basin deposits. It is suggested that these basin deposits inhibit vertical fluid flow and prevent the formation of a BSR, whereas the hydraulic conductivity of accreted sediments is sufficiently high to allow for pervasive gas migration. An elevation of the BSR beneath the flanks of a topographic high is interpreted as an indicator for local upflow of warm fluids along permeable pathways within outcropping accreted sediments.

At or directly beneath the seafloor, Parasound data reveal discontinuous zones of high reflectivity, which may indicate local cementation of surface sediments. In combination with GI-Gun data, the occurrence of these reflective areas can be related to the location of slope sedimentary basins acting as hydraulic seals. It is proposed that the seals
sometimes fail along faults extending beneath the BSR, leading to focused upflow of methane bearing fluid and the formation of carbonate pavements at the seafloor.

In general, information from seismic data about fluid flow may either be revealed by imaging flow-related changes of sediment physical properties, e.g. due to chemical alteration or early diagenesis, or by detecting and mapping potential fluid pathways as faults or high porosity zones. In any case, seismic data have to be combined with geological, hydrologic, or chemical data and need to provide sufficient quality and resolution.

The quality of seismic data is examined in the vicinity of the buried basement ridge by correlating seismic records with synthetic seismograms, which were calculated from ODP Leg 168 core logging data. At source frequencies above 100 Hz, seismic reflections can be related to variations in GRAPE density, indicating that density rather than seismic velocity is dominating sub-bottom variations in seismic impedance. Furthermore, GRAPE density is also sufficient to model the frequency-dependent amplitude decrease above the buried basement ridge. This confirms that low reflection amplitudes reflect lateral changes of in-situ physical properties and can not be assigned to processing artifacts or rough topography. Apart from providing ground truth for seismic data, a combined interpretation of seismic records and borehole data suggests that two different sedimentation regimes are separated by the buried basement ridge and that relative accumulation rates have changed during the depositional history of the area.

The correlation of synthetic seismograms with seismic records also indicates, that the quality of seismic data strongly depends on source frequency. At frequencies below 100 Hz, only few seismic reflections are reproduced from variations in GRAPE density, and their corresponding reflection time does not appear to be reliable. It can therefore be assumed that standard seismic data, which are often acquired using source frequencies below 50 Hz, would not be sufficient to be interpreted with respect to lithology and would be less sensitive to variations of physical properties than the high resolution data presented in this study.

The cost of increasing resolution is a decrease in signal penetration. To overcome this difficulty, seismic systems with different source frequencies were combined to provide maximum temporal resolution at each depth level. Consequently, some features can be

observed in more than one data set, providing a more reliable basis for seismic interpretation. Furthermore, a joint analysis of multi-frequency data reveals more information than can be provided by only one single data set. As an example, a combination of Parasound data (4000 Hz) and GI-Gun data (50-150 or 100-500 Hz) reveals that seafloor cements are related to the presence of faults (Cascadia margin) and that high porosity zones are located above basement highs (buried basement ridge). Water gun data (200-1600 Hz) further allow to trace high porosity zones throughout the sediment column and provide a link between observations at the seafloor and at greater depth. The frequency dependence of the amplitude decrease at high porosity zones, combined with information from Biot-Stoll's model, may indicate how porosity is related to permeability. In addition, the amplitude of the BSR at the northern Cascadia accretionary prism is also frequency dependent, suggesting that the transition between hydrate cemented sediment and the gas layer below is associated with a gradual change in seismic velocity.

In this and most previous studies in which dewatering processes were investigated, some questions have remained, including the relative importance of focused flow versus diffuse flow, the magnitude of flow, and the modulation of flow rates over time. A quantitative estimate of diffuse flow is very difficult since large volumes of fluid and sediment are involved and flow related changes of sediment physical properties, which could be detected and mapped by seismic data, may be small or gradual. Comprehensive studies of the formation, distribution, and thickness of gas hydrate layers may play a key role in future investigations on diffuse flow at convergent margins. On the other hand, a quantitative estimate of rates of confined flow and mass exchange requires numerical modeling, which is beyond the scope of this study. However, before flow can successfully be modeled, the thermal and hydraulic parameters as well as the nature and distribution of flow have to be determined. Conceptual models for the different fluid regimes, which can be used to constrain future models and investigations, were provided by high resolution seismic surveys during Cruise SO111. As an example, the thickness of highly reflective surface zones at the Cascadia margin as well as the heat flow anomaly, which is indicated by a local BSR elevation and has not yet been dissipated by thermal diffusion, may be indicative for the duration and intensity of flow. Furthermore, a new R/V Sonne cruise in combination

with an American cruise, both scheduled for late summer 2000, will build on the success of Cruise SO111. Above the buried basement ridge, the local patterns of fluid discharge will be imaged in even greater detail, and a quantification of local variations of sediment physical properties may be achieved by coring and supplementary measurements of porosity, density, and heat flow. Thus, the collection of high resolution multi-frequency seismic data can be considered as a step forward towards imaging the spatial extent of fluid flow and providing a basis for a subsequent quantification of flow-related processes.

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Maps in this thesis were created using Generic Mapping Tools (GMT) [Wessel and Smith, 1991]. Standard seismic processing was carried out with the Seismic Unix (SU) software package [Stockwell, 1997]. My own software for geometry processing incorporates free codes by Inge Nesbo for calculating accurate geographic positions as well as a subroutine available from the National Geophysical Data Center in Boulder, Colorado, for estimating magnetic declination. Biot-Stoll's model was implemented by Monika Breitzke. The latest version of the software for computing reflection coefficients using the state space approach was provided by Volkhard Spieß.

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