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Joint inversion of multichannel seismic reflection and wide-angle seismic data: Improved imaging and refined velocity model of the crustal structure of the north Ecuador-south Colombia convergent margin

W. Agudelo,^{1,2} A. Ribodetti,¹ J.-Y. Collot,¹ and S. Operto¹

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[1] Improving seismic imaging of the crust is essential for understanding the structural factors controlling subduction zones processes. We developed a processing work flow based on the combined analysis of multichannel seismic reflection (MCS) and wide angle (WA) reflection/refraction data to derive both shallow and deep velocities suitable for prestack depth migration and to construct a blocky velocity model integrating all identifiable seismic phases contained in MCS and WA data. We apply this strategy to the study of the north Ecuador-SW Colombia subduction margin to improve the imaging and geostructural interpretation of a splay fault and surrounding outer and inner margin wedges. Results show improvements over tomographic inversion of WA data only, such as (1) sediment velocity variation across the trench and margin slope that correlates with lateral lithologic changes, tectonic compaction and effect of mass wasting processes; (2) a two-layer velocity structure of the inner wedge basement that is consistent with the crust of an oceanic plateau; (3) a complex velocity structure of the outer wedge basement that consists of a deep, high-velocity $(5.0-5.5 \text{ km s}^{-1})$ core and a low-velocity zone $(3.8-5.0 \text{ km s}^{-1})$ associated with the major splay fault; (4) a \sim 1.3-km-thick, low-velocity (3.5–4.0 km s⁻¹) subduction channel that extends beneath the margin outer wedge. Both the splay fault and subduction channel are expected to direct fluid flows; and (5) downdip velocity increase $(5-6 \text{ km s}^{-1})$ in the subducting oceanic crust associated with a low (7.8 km s⁻¹) upper mantle velocity, possibly reflecting changes in rock nature or properties.

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1. Introduction

[2] Investigations of the shallow subduction plate interface that periodically release elastic strain during large earthquakes [*Hyndman et al.*, 1997] include controlledsource seismic imaging of the fault surface and adjacent structures to reveal physical features and changes in seismic reflectivity and to interpret processes associated with earthquake ruptures. Multichannel seismic reflection (MCS) and wide-angle seismic data are commonly used to illuminate acoustically active margin structures, including accretionary wedges [*Westbrook*, 1982], backstop structures [*Christeson et al.*, 2003; *Flueh et al.*, 1998], splay faults that branch on the interplate fault [*Park et al.*, 2002; *Nakanishi et al.*,

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2008], and the subduction channel [*Calahorrano et al.*, 2008] that is a fluid-rich, low-velocity sedimentary layer that is conveyed by and structurally squeezed between lower and upper plates [*Shreve and Cloos*, 1986].

[3] Prestack depth migration (PSDM) allows obtaining depth images of complex crustal structures from MCS data as well as 2-D seismic velocity model down to a depth that is controlled by the maximum source-receiver offset as well as by the signal-to-noise ratio and dip of reflections and the dominant source frequency [Ross, 1994]. Accurate seismic velocity modeling plays a central role in PSDM of seismic data and in their geological interpretation since it controls the accuracy of the positioning in depth of the reflectors and the quality of the stacking of the redundant information provided by multifold data set. In the last decade, migration velocity analysis (MVA) techniques have been developed to build accurate velocity models and depth images of structurally complex structures [Deregowski, 1990; Bleistein and Liu, 1992; Stork, 1992; Liu and Bleistein, 1995]. MVA and PSDM are performed iteratively until reflectors are flat on common image gathers.

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Figure 1. Location of seismic experiment. Black circles indicate the position of ocean bottom seismometers from UMR Géosciences Azur and ocean bottom hydrophones from GEOMAR. Crosses indicate the first (NW) and last (SE) shots for wide-angle (WA) acquisition profiles. Triangles show the position of first (SE) and last (NW) shots of the coincident multichannel seismic reflection (MCS) line SIS-44. Dashed line correspond to the WA segment of the profile. Heavy line indicates the combined (WA and MCS) acquisition segment of profile. Detailed bathymetry (100 m contours) is from *Collot et al.* [2005]. Black arrow is the Nazca–South America relative plate motion after *Trenkamp et al.* [2002]. Inset shows the rupture zone of 1906 great subduction earthquake.

[4] At oil exploration depths (less than 7 km), velocity information is obtained almost entirely from MCS data, further constrained by logging measurements. However, acquisition parameters commonly used in MCS surveys (offset \sim 5 km) represent an inherent limitation to its ability to estimate velocities at deeper levels [Lynn and Deregowski, 1981]. For example, one of the main problems of deep velocity modeling is the velocity-depth ambiguity, where the observed arrival time of a reflector can be equally well explained by a change in depth or a change in velocity. The error induced by this ambiguity has been related to the offset/depth ratio and the dominant frequency by O'Brien and Lerche [1988], Lines [1993], Ross [1994], and Rathor [1997], who have shown that a depth equivalent to the maximum offset marks the transition from a shallow level, where velocity-depth ambiguity error grows linearly with depth, and a deep level, where this error grows as the square of depth. Thus, what we will mean henceforth by deep level is the zone located deeper than a depth equivalent to the maximum offset of the acquisition geometry and where

errors in velocity and depth are significant. For example, most of the major geological structures at convergent margins, and notably the interplate contact, range within 5–30 km in depth [*Tichelaar and Ruff*, 1993; *Hyndman*, 2004], clearly in the deep level for common MCS acquisition parameters. Therefore constraining velocity distribution and depth of structures at convergent margins is severely affected by the limitations of MCS acquisition parameters. *Calahorrano et al.* [2008] recently used PSDM to produce an accurate velocity model across the Ecuador subduction zone and quantify physical property variations of sediment in the subduction channel down to a maximum depth of ~7 km.

[5] One way to ameliorate deep reflector imaging is to increase the offset/depth ratio. For example, ordinary offsets in wide-angle (WA) reflection/refraction acquisition geometry are in the order of tens of kilometers. Receivers at larger offsets would record information contained in critical and postcritical arrivals. PSDM methods for WA acquisition geometry have been developed and successfully applied in different geological contexts [*Lafond and Levander*, 1995;



Figure 2. MCS line SIS-44 across outer wedge basement and fore-arc basin after *Collot et al.* [2004]. (top) Line drawing. Note that fore-arc basin is more than 60 km wide and is dominantly undeformed, with the exception of a graben (G) associated with the outer wedge basement and an underlying crustal splay fault (SF). (bottom) A zoom showing time-migrated data across graben (G) and deformed basement (B) associated with the splay fault. U4 and U5 are major unconformities [*Marcaillou and Collot*, 2008]. Décollement (De) and top of oceanic crust (To). Mu are multiples. CDP, common depth point.

Simon et al., 1996; Zelt et al., 1998; Buske, 1999; Funck et al., 2001; Dessa et al., 2004]. However, in these PSDM methods, the migrated image resolution is limited by the receiver spacing, which, with the exception of some pioneer experiments [Dessa et al., 2004], continues to be relatively sparse (\sim 5 km). Thus, on one hand, high-resolution MCS acquisition geometry fails to accurately image deep level structures, and on the other hand, low-resolution WA acquisition geometry provides good constraints on deep level structures.

[6] In this study, we propose an integrated processing work flow applied to coincident MCS and WA data sets to develop a PSDM image and a crustal-scale blocky velocity model. We first combine the deep velocity structure inferred from first-arrival traveltime tomography of WA data with the shallow velocity model inferred from MVA to build a mixed velocity macromodel suitable for PSDM in order to perform the imaging of the full crustal structure. Then we integrate the structural information interpreted on the PSDM image, with the mixed velocity macromodel to build an initial blocky velocity model. This model is further refined by joint inversion of normal incidence traveltimes picked on stack sections, and refraction/wide-angle reflection traveltimes picked on ocean bottom seismometer (OBS) gathers. In the end, the PSDM image can be used in conjunction with the optimal blocky velocity model for the geological interpretation.

[7] This processing work flow is applied to a seismic transect consisting of coincident MCS and ocean bottom seismometer (OBS) seismic data collected across the north Ecuador–SW Colombia convergent margin (Figure 1) where large and great subduction earthquakes have occurred during the 20th century [*Kanamori and McNally*, 1982]. Using the time-migrated seismic section of the MCS line, *Collot et al.* [2004] interpreted a landward dipping, crustal reflector as a splay fault (splay fault interface, SF, in Figure 2) beneath the

inner trench slope break. However, this zone is structurally complex and difficult to image because of the following:

[8] 1. It is located in an area where the top of the basement has a high reflectivity, thus producing reverberations that interfere with primary reflections. Additionally, these reverberations have normal move out (NMO) velocities close to those of the primary reflectors, so that classical wave number-frequency domain techniques are not easily applicable to enhance the primary reflections.

[9] 2. The top of the basement shows an upward convexity (B in Figure 2) that tends to scatter seismic energy.

[10] 3. The 5–13 km depth and 25° dip place the interpreted splay fault in the zone of low spatial resolution for the MCS acquisition geometry [*Lambaré et al.*, 2003].

[11] Consequently, only one segment (reflector SF in Figure 2) of this splay fault was clearly identified on the time section. Applying the proposed processing work flow allows clarification of the complex structures of the north Ecuador–SW Colombia convergent margin.

[12] In this paper, we first model and interpret first and secondary arrivals picked in the ocean bottom seismometer/ ocean bottom hydrophone (OBS/OBH) gathers. We thus present our processing sequence, show how the PSDM image resolution improved, and discuss the uncertainties of the optimal blocky velocity model, prior to describing and discussing the main structures evident in the final PSDM image and the optimal blocky velocity model. Detailed geological interpretations and geodynamic implications of the velocity model and associated PSDM image are discussed by *Collot et al.* [2008].

2. Geodynamic Setting

[13] In Ecuador and Colombia, the margin basement consists of cretaceous oceanic and island arc rocks outcropping onshore [Goosens and Rose, 1973; Feininger and Bristow, 1980], on Gorgona Island (Figure 1 inset) [Dietrich et al., 1981; Kerr et al., 1996], and encountered in coastal drill sites in both Colombia [Salazar and Govea, 1974] and northern Ecuador [Evans and Wittaker, 1982]. These rocks are associated with oceanic terranes, including the Piñon formation in coastal Ecuador [Henderson, 1979; Feininger and Bristow, 1980], the Dagua terrane in the western cordillera of Colombia [McCourt et al., 1984; Cediel et al., 2003] and the Gorgona terrane mainly interpreted offshore beneath the Manglares fore-arc basin [Cediel et al., 2003; Kerr et al., 1996]. The terranes, which may belong to the fragmented Colombian-Caribbean Oceanic Plateau (CCOP) [Jaillard et al., 2009; Luzieux et al., 2006], were emplaced against South America during Late Cretaceous to early Cenozoic convergence of the Farallon-Nazca plate [Lonsdale and Klitgord, 1978; Cediel et al., 2003]. The Ecuador-Colombia margin is underthrust obliquely, eastward at 5.4 cm a^{-1} [*Trenkamp et al.*, 2002] by the morphologically complex Nazca plate, which includes the thick buoyant Carnegie Ridge [Sallares and Charvis, 2003] and the Yaquina Graben [Lonsdale and Klitgord, 1978] (Figure 1). Swath bathymetry acquired during the Amadeus project [Collot et al., 2005] shows the southern Colombia trench to be partially filled by a deep-sea fan associated with the Esmeraldas canyon (Figure 1). In this region, although no accretionary wedge was interpreted, a crustal splay fault that

branches from the plate interface separates the outer and inner margin wedges (Figure 2) [Collot et al., 2004; Marcaillou et al., 2006]. The inner margin wedge comprises the 60-km-wide, poorly deformed Manglares fore-arc basin [Marcaillou and Collot, 2008]. In 1906 a remarkable Mw 8.8 subduction earthquake ruptured the north Ecuador-SW Colombia margin segment from the northern flank of the Carnegie Ridge to Buenaventura in Colombia (Figure 1 insert) [Kanamori and McNally, 1982]. This region was ruptured again in 1942, 1958, and 1979 by magnitude 7.7 to 8.2 subduction earthquakes [Mendoza and Dewey, 1984; Swenson and Beck, 1996]. The seismic transect presented in this paper is intended to show details of the crustal structures of this highly seismogenic north Ecuador-south Colombia margin. Figure 3 shows the final velocity model developed in this paper, which highlights the main structural units revealed by the joint processing of the MCS and WA data.

3. Data Set Acquisition and Processing

[14] The MCS and WA seismic data that were collected independently along a geophysical transect across the northern Ecuador–SW Colombia convergent margin (Figure 1) are analyzed.

3.1. MCS Data

[15] The MCS line SIS-44 was collected during the SISTEUR cruise [Collot et al., 2002] with a 4.5-km-long, 348-channel streamer. A 45-L air gun seismic source fired every 50 m provided a 43-fold data set. MCS data were processed using the Geovecteur (2) seismic processing software, following a preprocessing flow designed to prepare data for PSDM. Preprocessing included attenuation (0.001 factor) of the noised traces amplitudes, a band-pass filtering (3, 6, 50, 60 Hz), multiple attenuation in the frequencywave number (FK) domain, multiple attenuation using radon transform, minimum phase conversion of the signal, predictive deconvolution, and a loose external mute. Then, this preprocessed data set was depth migrated using the PSDM method described in the section 5.2 (prestack depth migration (PSDM) and migration velocity analysis using only MCS information). The final PSDM images will be shown after application of a depth variant automatic gain control (AGC) of amplitudes.

3.2. WA Data

[16] A coincident WA line was collected during the SALIERI cruise [*Flueh et al.*, 2001]. Shots were produced at a constant time interval of one minute giving an average shot spacing of 150 m using a 128 L air gun source and recorded by 10 ocean bottom hydrophones (OBH) from GEOMAR (Kiel, Germany) and 12 ocean bottom seismometers (OBS) from GEOAZUR (UMR Géosciences Azur, Villefranche-sur-Mer, France). Although WA data quality is variable, it is generally good with some arrivals being clearly identified at 120 km offsets in some record sections. Processing on board consisted in correcting for clock drift during deployment and inverting the direct arrivals to obtain the instrument location (OBH and OBS) and orientation (OBS) [*Christeson*, 1995]. Further processing included the application of a Butterworth filter (low cut of 5 Hz, high cut



Figure 3. Optimal blocky velocity model developed by the joint analysis of the MCS and WA data. Black lines are layer boundaries and white lines are isovelocity contours in km s⁻¹. SD, sedimentary cover; IWB1, upper inner wedge basement; IWB2, lower inner wedge basement; OWB, outer wedge basement; SC, subduction channel; OCA, oceanic crust layer A; OCB, oceanic crust layer B.

of 15 Hz), predictive deconvolution (whitening) and amplitude equalization.

4. Modeling and Interpretation of WA Seismic Data

[17] First and secondary arrivals were picked for all receivers and are listed in Table 1. Arrivals are displayed for selected OBHs 113 and 117 and OBSs 121, 123, 125, and 129 representative of groups of receivers (Figures 4–9). To facilitate the interpretation of the main arrivals, they were modeled in the velocity model of Figure 3 using a 2-D forward modeling technique [*Zelt*, 1999].

4.1. Arrivals P1 and R1, Sedimentary Cover

[18] In the westernmost OBHs (107–112) (Figure 1), sediment thickness and velocities are too low to produce a P1 refracted arrival that differentiates from the direct wave $(\sim 1.6 \text{ km s}^{-1})$. From OBHs 113 to 118 (OBHs 113 and 117, Figures 4 and 5), the refracted arrival P1 in the deformation front is easily distinguished from the direct wave with a higher apparent velocity of ~ 2.0 km s⁻¹. For OBSs 120 to 130, located on the margin (Figure 1), P1 arrival breaks closer to the direct arrival but presents a slight curvature associated with a positive velocity vertical gradient that gradually separates it from the direct arrival (OBSs 121 and 125, Figures 6 and 8). In the lower part of the sedimentary layer apparent velocities reach 3.0 km s⁻¹ on OBS 123 (Figure 7). The reflection from the margin basement R1 is easily identified in the records of 15 OBH/OBSs (for example, OBH 113 in Figure 4 and OBSs 121, 123, and 129 in Figures 6, 7, and 9). In OBHs 115 and 118 and OBS 120, this arrival is not identified, likely because beneath these OBH/OBS, a highly deformed basement generates seismic scattering, as shown by the MCS time-migrated section (Figure 2).

4.2. Arrivals P2 and R3, Outer Wedge Basement

[19] A clear refraction arrival (P2) is observed in OBHs 115-120 (example in OBH 117, Figure 5), and in OBSs 121 (Figure 6) and 122. P2 average apparent velocity is 3.0 km s⁻¹ (Table 1). reflection R3 from the top of the oceanic crust is observed in OBHs 116 and 117 (Figure 5) and OBSs 120 and 121 (Figure 6).

4.3. Arrivals Pg1 and Rf, Upper Inner Wedge Basement

[20] Refracted arrival Pg1 is easily identified as a large slope break in record sections for OBS 118, OBH 120 and OBSs 121-130 (for example, OBS 121 in Figure 6, OBS 123 in Figure 7, OBS 125 in Figure 8, and OBS 129 in Figure 9) and was used to model velocities in the margin basement. In OBH 120 and OBS 121 (Figure 6) and OBS 122, Pg1 seaward branches and landward branches are highly asymmetric indicating that velocity in the westernmost part of the layer (apparent velocity $\sim 3.0 \text{ km s}^{-1}$) are lower than eastward (apparent velocity $\sim 4.5 \text{ km s}^{-1}$). In the easternmost part of the WA line, Pg1 apparent velocity reaches ~ 6.0 km s⁻¹ (OBS 129 in Figure 9). A weak WA reflection Rf from the interface between the upper and lower margin basement is locally identified in OBSs 120-128 (for example, OBS 121, Figure 6; OBS 123, Figure 7; and OBS 125, Figure 8).

4.4. Arrival Pg2 and R2, Lower Inner Wedge Basement

[21] A gradual slope change marks the separation between refracted arrivals Pg1 and Pg2. While Pg1 has apparent velocities of ~5.0 km s⁻¹, Pg2 apparent velocities are >6.0 km s⁻¹ (OBS 121, Figure 6, OBS 123, Figure 7, and OBS 125, Figure 8). Phase R2 reflected from the splay fault SF (Figure 3) and its connection to the interplate



Figure 4. Record section and raypaths for OBH 113. (a) Ray tracing through the optimal blocky velocity model (Figure 3). (b) Record section. Data processing included time- and offset-dependent frequency filtering and predictive deconvolution. Time axis is reduced by 6 km s^{-1} . Modeled traveltimes (yellow lines) are superimposed on the data and on the picked traveltimes (red lines with error bars). Labels of the interpreted phases are plotted (see Table 1). P1, refraction from the sedimentary cover of the margin; R1, reflection from the base of the sedimentary cover; Pc, refracted from the downgoing oceanic crust; PmP, reflected from the Moho: Pn, head wave along Moho. (c) Observed data.



Figure 5. Record section and raypaths for OBH 117. (a) Ray tracing in the optimal blocky model (Figure 3). (b) Record section with picked (red lines with error bars) and modeled traveltimes (yellow lines), and labels of the interpreted phases (see Table 1). The Psc phase is a refracted wave at the front of the subduction channel with an apparent velocity \sim 4.0 km s⁻¹ and is identified only on this OBH. P1, refraction from the sedimentary cover of the margin; R1, reflection from the base of the sedimentary cover; R3, reflection from the top of oceanic crust; P2, refraction from the frontal block; Pc, refracted from the downgoing oceanic crust; PmP, reflected from the Moho: Pn, head wave along Moho; SCSZ, subduction channel shadow zone. (c) Observed data.



Figure 6. Record section and raypaths for OBS 121. (a) Ray tracing in the optimal blocky model (Figure 3). (b) Record section (vertical component) with picked (red lines with error bars) and modeled traveltimes (yellow lines), and labels of the interpreted phases (see Table 1). P1, refraction from the sedimentary cover of the margin; R1, reflection from the base of the sedimentary cover; R2, reflection on upper/lower margin basement interface; P2, refraction from the frontal block; Pg1, refraction from the upper inner wedge basement; Pg2,: refraction from the lower inner wedge basement; Pg2,: refraction from the Moho: Pn, head wave along Moho; SCSZ, Subduction channel shadow zone; SFSZ, splay fault shadow zone. (c) Observed data.

contact is interpreted in OBSs 121–125 (Figures 6, 7, and 8) and 129 (Figure 9).

4.5. Arrival Pc, Oceanic Layer A and B

[22] Arrival Pc with apparent velocity ~ 6.5 km s⁻¹ is clearly identified in OBHs 107–114 (for example, OBH 113, Figure 4), located in the trench zone. In records of OBHs 116–120 (for example, OBH 117, Figure 5) and

OBSs 121–130 (for example, OBSs 121, Figure 6; 123, Figure 7; and 125, Figure 8) Pc arrival is strongly attenuated, indicating a shadow zone [*Gailler et al.*, 2007].

4.6. Arrivals PmP and Pn, Moho and Mantle

[23] PmP arrival is identified and successfully modeled in all OBS/OBS (Figures 4–9) but OBH120. Pn arrival is





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observed and modeled in all OBS/OBH (Figures 4-9) with the exception of OBH 115, 118 and 120.

4.7. Shadow Zones

[24] The single shadow zone (SZ) identified by *Gailler et* al. [2007] appears to be dual on OBS 122-124 (for example, OBS 123, Figure 7): (1) a strong attenuation of phase Pg1 at offset ~25 km in OBS 123 (Figure 7) is indicative of a low-velocity zone (splay fault shadow zone, SFSZ) associated with the splay fault and outer wedge basement rocks (outer wedge basement, OWB, in Figure 3), and (2) refracted phase P2 also shows attenuation near offset \sim 15 km (OBS 123, Figure 7) and is not continuous with phase Pc, supporting the existence of a shadow zone (SCSZ) associated with the subduction channel (SC) and the top of the underlying oceanic crust (OCA in Figure 3). These two shadow zones cannot be differentiated in OBS records 124-130 (Figures 8 and 9), where a single shadow zone (SZ) is interpreted. The subduction channel shadow zone can be extended downdip by a low-velocity zone resulting from the velocity contrast between the base of the IWB2 and OCA layer near the interplate contact, for x > 40 km (Figure 3).

[25] The quality of the optimal blocky model (Figure 3) can be assessed qualitatively in Figures 4–9 and quantitatively by using the fit between predicted and observed traveltimes summarized for each picked phases in Table 2. The total RMS-traveltime misfit in the optimal blocky velocity model is 0.150 s (Table 2), which is within the 0.109–0.169 RMS range of the different blocky models used during the joint inversion. The RMS increase results from the numerical limitation of the Zelt software [*Zelt*, 1999] when modeling traveltimes in very complex media, such as our model, the increased geometrical complexity of which results from the increasing number of observed phases progressively introduced during the joint inversion process described in sections 5.6 and 5.7.

5. Seismic Processing Work Flow

5.1. Overall Description

[26] The aim of the processing work flow (Figure 10) is to design a joint analysis of the MCS and WA data in order to produce a depth-migrated section and a blocky velocity model which integrates as much as possible the information provided by the two data sets.

[27] The seismic work flow is subdivided in five main steps which are described hereafter (Figure 10): the first step is the PSDM of the MCS data using a macromodel developed by MVA only. This first step should provide a good image of the shallow structure but cannot guarantee reliable imaging of the deeper structure due to missing constraint on the deep velocities in the macromodel. The second independent step consists of first-arrival traveltime tomography applied to the WA data to develop a smooth velocity model of the crustal structure In the third step, we mix the macromodel derived from MVA, with the smooth velocity model developed by first-arrival traveltime tomography to build a new macromodel for PSDM. This macromodel improves the imaging of the deep reflectors while preserving the quality of the imaging of the shallower ones. The fourth and fifth steps contribute to the development of the optimal blocky velocity model. They integrate the structural information interpreted on the PSDM images, the normal incidence traveltimes picked on the time stack section and the refraction/wide-angle reflection traveltimes picked on the OBS gathers. The optimal blocky velocity model (Figure 3) provides a quantitative structural model parameterized by P wave velocities, which integrates all the information coming from the MCS data and the traveltimes from the WA data.

5.2. Prestack Depth Migration and Migration Velocity Analysis Using Only MCS Information

[28] An initial macromodel for PSDM was inferred from NMO velocities after conversion into interval velocities using the Dix formula (see step labeled 1 in Figure 10). A 2-D velocity model was interpolated from the resultant velocitydepth profiles and subsequently smoothed (Figure 11a) for PSDM (Figure 11b). To migrate the data, we used the true amplitude ray plus Born PSDM method [Lambaré et al., 1992; Thierry et al., 1999] based on asymptotic Green functions computed by dynamic ray tracing [Lambaré et al., 1996]. Although this method theoretically returns a quantitative information on the reflectors corresponding to velocity perturbations at the interfaces, we will make no attempt in this paper to analyze such information. We will focus on the assessment of the accuracy of the migrated image in terms of focusing and positioning in depth of the reflectors. This appraisal can be performed thanks to common image gathers (CIG) which are collections of depthmigrated traces for a fixed horizontal distance x and for all diffraction angles θ (or offsets) [Xu, 2001] (Figure 12). The redundant image of the reflectors for each offset or diffraction angle in the CIGs must be flat to guarantee their optimal subsequent stack. Indeed, assemblage of all the stacked CIGs in the distance-depth domain provides the final PSDM image. Iterative flattening of reflectors in CIGs is the basis of MVA methods [Deregowski, 1990; Bleistein and Liu, 1992; Stork, 1992; Liu and Bleistein, 1995]. In this paper, we implemented the MVA method proposed by Al-Yahya

Figure 7. Record section and raypaths for OBS 123. (a) Ray tracing in the optimal blocky model (Figure 3). (b) Record section (vertical component) with picked (red lines with error bars) and modeled traveltimes (yellow lines), and labels of the interpreted phases (see Table 1). P1, refraction from the sedimentary cover of the margin; R1, reflection from the base of the sedimentary cover; Pg1, refraction from the upper Inner Wedge basement; Pg2, refraction from the lower inner wedge basement; R2, reflection from the Splay fault and plate interface; Pc, refracted from the downgoing oceanic crust; PmP, reflected from the base of the sedimentary cover; R2, reflection from the base of the sedimentary cover; R2, reflection from the base of the sedimentary cover; R2, reflection from the Splay fault and plate interface; P2, refraction from the frontal block; Pg1, refraction on upper/lower margin basement interface; P2, refraction from the frontal block; Pg1, refraction from the upper Inner Wedge basement; Pg2, refraction from the lower inner wedge basement; Pg2, refraction from the sedimentary cover of the margin; R1, reflection from the base of the sedimentary cover; R2, reflection from the Splay fault and plate interface; R3, reflection from the top of oceanic crust; Rf, reflection on upper/lower margin basement interface; P2, refraction from the frontal block; Pg1, refraction from the upper Inner Wedge basement; Pg2, refraction from the lower Inner Wedge basement; Pc, refracted from the downgoing oceanic crust; PmP, reflected from the Moho; Pn, head wave along Moho; SCSZ, subduction channel shadow zone; SFSZ, Splay fault shadow zone. (c) Observed data.



Figure 8. Record section and raypaths for OBS 125. (a) Ray tracing in the optimal blocky model (Figure 3). (b) Record section (vertical component) with picked (red lines with error bars)and modeled traveltimes (yellow lines), and labels of the interpreted phases (see Table 1). P1, refraction from the sedimentary cover; R2, reflection from the base of the sedimentary cover; R2, reflection from the splay fault and plate interface; Rf, reflection on upper/lower margin basement interface; P2, refraction from the frontal block; Pg1, refraction from the upper inner wedge basement; Pg2, refraction from the lower inner wedge basement; PmP, reflected from the Moho; Pn, head wave along Moho; SZ, shadow zone. (c) Observed data.



Figure 9. Record section and raypaths for OBS 129. (a) Ray tracing in the optimal blocky model (Figure 3). (b) Record section (vertical component) with picked (red lines with error bars) and modeled traveltimes (yellow lines), and labels of the interpreted phases (see Table 1). P1, refraction from the sedimentary cover of the margin; R1, reflection from the base of the sedimentary cover; R2, reflection from the splay fault and plate interface; Pg1, refraction from the upper inner wedge basement; Pg2, refraction from the lower inner wedge basement; Pc, refracted from the downgoing oceanic crust; PmP, reflected from the Moho; Pn, head wave along Moho; SZ, shadow zone. (c) Observed data.

[1989], which is conceptually simple and robust. During MVA, the velocity macromodel is iteratively corrected, until the CIGs panel are sufficiently flattened. After 2 iterations of PSDM plus MVA, we obtained an updated velocity model (v_{mcs}) for the shallower part of the seismic section (Figure 11c). Errors in velocity estimation grow quickly for depths greater than the maximum offset of the streamer (for MCS data is 4.5 km) [*Lines*, 1993; *Ross*, 1994].

5.3. Processing of WA First-Arrival Traveltimes

[29] A smooth two-dimensional velocity model along the same geophysical transect was developed from the WA data by first-arrival traveltime tomography using the method of *Korenaga et al.* [2000] (step labeled 2 in Figure 10). The reader is referred to *Gailler et al.* [2007] for a detailed description of this step of the processing work flow. The

Table 1. Classification of Major Seismic Phases for Section SIS-44

Phases Type		Apparent Velocity (km s ⁻¹)	Interpretation		
P1	Refracted	$\sim 2.0 - 3.0$	Sedimentary cover		
Pg1	Refracted	$\sim \!\! 4.5$	Upper inner wedge basement of the margin		
Pg2	Refracted	~5.5	Lower inner wedge basement of the margin		
P2	Refracted	~ 3.0	Frontal block		
Pc	Refracted	~ 6.5	Oceanic crust		
Pn	Head wave	~ 7.0	Along Moho		
R1	Reflected	-	Sedimentary cover/oceanic crust interface		
		-	Sedimentary cover/upper basement interface		
Rf	Reflected	-	Upper/lower margin basement interface		
R2	Reflected	-	Splay fault-interplate contact		
R3	Reflected	-	Top of oceanic crust beneath OWB/décollement		
PmP	Reflected	-	Reflected from the Moho		

velocity model developed by *Gailler et al.* [2007] is shown in Figure 11d.

5.4. Construction of the Mixed Velocity Model for PSDM

[30] We build a composite velocity model by mixing the velocity model inferred from PSDM plus MVA (hereinafter referred to as MCS model) (Figure 11c) and that inferred from first-arrival traveltime tomography (hereinafter referred to as WA model) (Figure 11d) (step labeled 3 in Figure 10). The resultant mixed model is composed of three zones (Figure 11e): (1) The shallow part corresponds to that of the MCS model, and the thickness of this zone is estimated from the maximum acquisition offset and reaches a value of 5 km. (2) The deep part corresponds to that of the WA model. (3) A transition part corresponds to a weighted average between the MCS and the WA models and allows avoidance of a sharp discontinuity between the shallow and deep parts of the mixed model. The mixed velocity model (Figure 11e) was used as velocity macromodel for PSDM of MCS data. The migrated image using the mixed velocity model is shown in Figure 11f and can be compared with that obtained with the initial macromodel inferred from the interval velocities (Figure 11b). Note that the transition zone of the mixed model was refined using one iteration of MVA plus PSDM (Figure 11f).

5.5. Accuracy and Improvements of the PSDM Image Using the Mixed Velocity Model

[31] Accuracy and validity of the improvements of the PSDM images from Figures 11b-11f can be assessed by analyzing CIG panels (Figure 12). Iterative flattening of reflectors in CIGs is performed during PSDM until CIG panel are sufficiently flat. After two iterations of PSDM plus MVA, we obtained an updated velocity model (Figure 11c) for the shallower part of the seismic section and CIG panels appear flattened for depths shallower than 4 km (CIG panels at iteration 2). When the mixed velocity model (Figure 11e) is used, the effect of the WA derived velocities on PSDM imaging can be evaluated more quantitatively by examination of the CIG panels (Figure 12 for the mixed model). At larger depths (z > 5 km), velocities inferred from NMO velocity analysis are up to ~ 2.0 km s⁻¹ lower than velocities derived from wide-angle data (compare Figures 11a and 11e). These velocity variations led to significant variations of the position in depth of the deep reflectors in the migrated images. As expected, we noted that, including WA

velocities in the mixed model resulted in wider, diffraction angle at larger depths, thus providing better constrains on the depth of the deep reflectors, and increasing their lateral coherency (see, for example, reflectors D, Rt, and To in Figure 12).

[32] Two zooms of Figures 11b and 11f are presented in Figures 13 and 14, respectively, to pinpoint changes and resolution improvements. Figure 13 is centered on the upper 5 km, whereas Figure 14 focuses on a deeper zone, centered on the splay fault and extending down to a 15-km depth. Major reflectors (U3, SF1) discussed by *Collot et al.* [2008] are labeled in Figures 13 and 14 to highlight improvements. One can note an improved continuity of the image of faults F1 and F2 associated with the summit graben in Figure 13.

[33] From top to bottom improvements include imaging of the unconformity U3 of likely late Oligocene-early Miocene age [Marcaillou and Collot, 2008] that gained in strength and continuity, and moved upward by $\sim 100-$ 150 m. A better continuity was obtained for reflector U4 of probable late Eocene age [Marcaillou and Collot, 2008] that was shifted up by $\sim 200-300$ m and can now be identified across most of the line. The group of reflectors labeled U5 and interpreted as the top of the margin basement [Marcaillou and Collot, 2008] has collapsed to form a thinner and better resolved reflector, especially under the Graben and the Canyon, where the top of U5 shifted upward by ~ 400 m. Although the resolution improved, a loss in the amplitude of U5 reflectors is observed. This amplitude loss results from the radon transform multiple elimination technique that we applied to the MCS data prior to PSDM with

Table 2. Number of Picked Traveltimes, RMS of Traveltime Misfits, and χ^2 Value for Interpreted Phases^a

Phase	Picked Traveltimes	RMS (s)	χ^2
P1	979	0.132	1.736
Pg1	2610	0.118	1.387
Pg2	899	0.116	1.355
P2	662	0.233	5.439
Pc	3076	0.138	1.928
Pn	5748	0.146	2.135
R1	823	0.107	1.136
Rf	743	0.218	4.757
R2	130	0.165	2.757
R3	640	0.203	4.112
PmP	1678	0.169	2.853
TOTAL	17988	0.150	2.256

^aThese traveltimes were computed in the optimal blocky velocity model of Figure 3.



Figure 10. Sketch of the processing flowchart of MCS and WA data. (1) The first step is iterative PSDM plus MVA to derive a shallow velocity model (MCS model) and an accurate image in depth of the shallow structure down to a maximum depth of the order of the streamer length (5 km). (2) First-arrival traveltime tomography is applied to WA data to develop a smooth velocity of the crustal structure down to depths of $\sim 10-20$ km (WA model). (3) A mixed velocity model is built by weighted averaging of the velocity model inferred from the two previous steps. The shallow and deep parts of the mixed model correspond to that of the MCS and WA models, respectively. A transition zone between the shallow and deep parts is a weighted average of the MCS and WA models. Migration of MCS data using the mixed model as macromodel provides the optimal PSDM image of both the shallow and deep structures. (4) An initial blocky velocity model is built from the structural information picked on the PSDM image and the velocities inferred from the MCS and WA models. (5) The blocky velocity model is further refined by joint inversion of normal incidence traveltimes and WA reflections. The PSDM image derived in step 3* as well as the optimal blocky velocity model (step 5*), are used for the final geostructural interpretation.

the mixed velocity model in order to attenuate interferences between U5 reflectors and water bottom multiples. Evidence for a major event dipping landward between 3.5 and 5 km depths (SF1 in Figure 13) is tentatively identified as the upward termination of splay fault SF1. The upward shift of unconformities U3 to U5 is consistent with the shallow sediment velocity decrease in the mixed velocity model (Figure 11e) relative to the initial velocity model (Figure 11a).

[34] Figure 14 shows improvements in reflector continuity and strength in the splay fault and subduction channel

regions. The band of reflector G may have moved up by \sim 800 m. Reflectors associated with splay faults SF1 and SF2 improved continuity, increased their dip, and shifted upward possibly by \sim 0.3–1.5 km. Reflector To which is interpreted as the top of the oceanic crust [*Collot et al.*, 2004] has gained in continuity and possibly shifted upward by \sim 1.0–1.5 km. The image of overlapping reflectors D interpreted as duplexes was clearly improved, and slightly (0.2 km) shifted downward and tilted landward, and reflector Rt, which is the subduction channel roof thrust [*Collot et al.*, 2008] was strengthened and apparently pull down by

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Figure 11. Steps used to obtain the optimal PSDM image along line SIS-44. (a) Velocity model from depth-converted, interpolated, and Gaussian-filtered ($\tau_x = 3000 \text{ m}$, $\tau_z = 400 \text{ m}$) normal move out velocities (NMO) for MCS data. (b) PSDM image using the initial NMO velocity model (ITERAT 1) presented in Figure 11a. (c) MCS Velocity model after 2 iterations of MVA. (d) WA velocity model from first-arrival traveltime tomography [*Gailler et al.*, 2007]. Note a low-velocity zone (box labeled LVZ) that coincides with the location of the splay fault around x = 40-45 km and z = 10 km. (e) Mixed velocity models. The upper (z < 5 km) and the lower deep (z > 10 km) zones correspond to that of the MCS and WA velocity models, respectively. The transition zone bounded by dashed lines ($5 \le z \le 10$ km) was obtained from a weighted sum (quadratic function) of the MCS and WA models to avoid a sharp discontinuity in the mixed velocity model. (f) Optimal PSDM image using the mixed velocity model (Figure 11e). Black arrows indicate reflectors improved during this process.

 \sim 0.3–0.5 km at the bottom right corner of Figure 14b. The downward shift of reflectors D and Rt in the eastern and laterally homogeneous part of the velocity model is consistent with the overall velocity increase in the margin rocks overlaying reflector D and Rt. In contrast, in the more complex

region of the model, between km 25 and 40, where both vertical and lateral heterogeneities were incorporated, reflectors such as G, SF1, SF2, and To may have shifted upward, although it might speculatively to tentatively correlate migrated dipping reflectors within a highly heterogeneous





Figure 13. (a) Zoom of a shallow section of the PSDM seismic section presented in Figure 11b. (b) Zoom of a shallow section of the PSDM seismic section plotted in Figure 11f. Arrows enable comparing major reflectors between Figures 13a and 13b. These reflectors are defined by *Collot et al.* [2008]: F1 and F2 are normal faults, U3 and U4 and U5 are unconformities, and SF1 is interpreted as the top part of the splay fault. Ellipses show zones where resolution of the migrated image improved. U5 dashed line is the top of the oceanic basement. Note that in Figure 13a, reflector U5 is thicker and interferes with the water bottom multiple; in Figure 13b, U5 moved up, for example, from 3.2-km depth to 2.8-km depth at offset x = 30 km. Resolution of U5 improved, although its amplitude decreased because we applied a multiple elimination filter before migration. The thin vertical dashed lines with white dots indicate the location of common image gathers shown in Figure 12.

velocity model. However, we observe that SF1, SF2, and To reflectors are better focused in the mixed image (Figure 14b) than in the initial one (Figure 14a), and we find a good coherency between the geometry of the well-focused migrated reflectors and the large-scale low-velocity zone associated with the SF region in the mixed velocity model (Figure 14c).

5.6. From a Smooth to a Blocky Velocity Model: Joint Refraction and Reflection Traveltime Inversion

[35] A blocky (i.e., layered) velocity model was built from joint inversion of refraction and reflection traveltimes (step labeled 4 in Figure 10). The reflected arrivals provide the necessary constraints on the position of the reflectors modeled as velocity discontinuities from where energy is reflected. One way to include these reflectors is the construction of a blocky model composed of velocity layers bounded by interfaces (see *Lailly and Sinoquet* [1996] for a general discussion on blocky versus smooth models for seismic imaging of complex geologic structures). We used the parameterization of *Zelt and Smith* [1992] that divides each layer in trapezium whose vertex are nodes that define either the interface geometry or the upper and lower layer velocities. Velocity within each trapezium is a bilinear interpolation of the values in the velocity nodes placed at its vertex.

[36] In the blocky model, four kinds of observed times can be inverted for the characterization of each layer: (1) shortoffset reflection times in MCS data (this kinematics information was previously processed by MVA to build the shallow part of the velocity model and to migrate the MCS data, and reflectors in the PSDM image can be picked and integrated in the blocky velocity model), (2) zero-offset reflection times that are picked on the stacked time section (these traveltimes can be modeled by computing normal incidence rays from each interface following the exploding reflector concept [*Wang and Braile*, 1996; *McCaughey and Singh*, 1997]); (3) wide-angle reflections, and (4) refraction

Figure 12. Common image gathers (CIG) at selected points (a) x = 29 km, (b) x = 34 km, (c) x = 39 km, (d) x = 41.5 km, and (e) x = 49 km, along line SIS-44. Iteration 1: CIGs computed with the initial velocity model inferred from NMO velocities; Iteration 2: CIGs computed with the corrected velocity model. Mixed: CIG panels corresponding to the migrated image obtained with mixed velocity model. CIG locations are shown in Figures 13 and 14. Note progressive flattening of reflectors (from top to bottom) during iterations and flatten reflectors (see SF1, SF2, D, Rt, and To) appear more clearly at depth >6 km for the mixed velocity model. Note zones where coherency of traces of the CIG panels increases and deep zones where energy is reinforced when mixed model is used. Horizontal red dashed lines indicate the transition zone used to build the mixed model. Zooms I and II show the details of the upper section (<2.8 km) of CIG km 41.5 and km 49, underlying reflector flattening between iterations 1 and 2. The flattening is preserved in CIG panels from the mixed model.



Figure 14. (a) Zoom of a deep section of PSDM seismic section presented in Figure 11b. (b) Zoom of a deep section of PSDM seismic section plotted in Figure 11f. Arrows are plotted on both Figures 14a and 14b to tentatively compare different reflectors: G is a band of strong reflections in the margin basement, To is top of the oceanic crust, Rt is roof thrust/top of subduction channel, SF1 and SF2 are splay faults, and D is duplex [*Collot et al.*, 2008]. Note that the resolution and the continuity of reflectors improved in Figure 14b in the zones associated with the subduction channel Sc and the splay faults SF1 and SF2. The thin vertical dashed lines with white dots indicate the location of common image gathers shown in Figure 12. Dashed lines are intended to show the lateral coherency of major reflectors representing To, Rt, SF1, and SF2. (c) Zoom of the mixed velocity model (Figure 11e) with superimposed line drawing of the well-focused migrated reflectors taken from Figure 14b.

traveltimes that are conventionally picked on OBS gathers (Table 2).

[37] The mixed velocity model is a smooth 2-D function. Transformation from a smooth to a blocky model is obtained as follows: main reflectors, that is, those that we interpreted as corresponding with wide-angle reflections, were picked in the depth-migrated image obtained from the mixed velocity model (Figure 15a), to define the initial geometry of layer interfaces (Figure 15b). As it was described before, in Zelt's parameterization (RAYINVR software [*Zelt* and Smith, 1992]), velocity within each layer is defined using only values in its upper and lower boundaries. With \sim 10 times less degrees of freedom, in general the blocky velocity model must not reproduce exactly the smooth velocity model and its RMS time misfit must be higher. There are several ways to extract velocities for each layer



Figure 15. Transforming the smooth mixed velocity model into a "blocky" model. (a) Picking of major reflectors (crosses) on the depth migrated image obtained by using the mixed velocity macromodel. (b) Blocky model obtained by setting the velocities of the mixed velocity model in each layers delineated by the interpretation of the migrated image in Figure 15a. (c) Updated blocky velocity model after first-arrival traveltime inversion. Comparison of the thin black lines (associated with the main reflectors) in Figures 15b and 15c highlight the evolution of the geometry.

top and bottom using the information of the smooth model. After several tests, we found that the approach that less increased the time misfit was to preserve the mean velocity gradient for each layer and afterward invert layer velocities to fit wide-angle reflection/refraction arrivals, holding the interface geometry fixed (Figure 15c). This layered velocity model was used as the starting model for the final step of the joint inversion presented hereafter.

5.7. Joint Inversion of Normal-Incidence Reflection and Refraction/Wide-Angle Reflection Traveltimes

[38] The last step of the blocky velocity model building consists of joint inversion of normal-incidence reflection and refraction/wide-angle reflection traveltimes to update the interface geometry and the velocity parameters of the blocky velocity model [*Wang and Braile*, 1996; *McCaughey and Singh*, 1997] (step labeled 5 in Figure 10). Computation and inversion of both normal-incidence and refraction/wide-angle reflection traveltimes were performed with the RSTTI software [*Operto*, 1996] which is an extension of the RAYINVR code [*Zelt and Smith*, 1992].

[39] The normal incidence traveltimes associated with the main reflectors (top of the oceanic crust To, top of the margin basement B, splay fault SF, and an intra basement reflector IB) were picked on the time stack section (Figure 16a) while refraction and wide-angle reflection traveltimes are picked on the OBS gathers. The full set of traveltimes was inverted simultaneously following a layer stripping approach to derive the final velocity model. Relative weight of WA and MCS data information during the combined inversion is controlled by the density of normal incidence rays. We choose a normal incidence ray density such that numbers of WA reflection/refraction rays and of MCS zero-offset (normal incidence) rays were roughly equal. Example of normalincidence ray tracing for the main reflectors of the blocky model is illustrated in Figure 16c. The agreement between the observed normal-incidence traveltimes with that computed in the final velocity model is shown in Figure 16e. The difference between the geometry of the reflectors of the starting and final velocity models is shown in Figure 16d. One can note that the deep reflectors are those which were mainly modified. This is consistent with the fact that the position in depth of these reflectors are poorly constrained by the MCS data in contrast with the traveltimes of the refractions and wide-angle reflection traveltimes.

[40] The optimal blocky velocity model (Figure 3) accounts for the both the normal-incidence and wide-angle reflection traveltimes along the main seismic reflectors such as the splay fault and the interplate boundary. Illustration of the WA reflection traveltime modeling from these 2 interfaces is provided in Figure 17. The blocky model was smoothed to tentatively perform an ultimate PSDM of line SIS-44. The resulting image was not significantly improved compared to the PSDM image (Figure 11f) obtained from the mixed velocity model, probably because the smoothing of the optimal blocky velocity model provides a smooth macromodel close to the mixed one. However, the optimal blocky model which integrates all the available seismic information is consistent with the PSDM image and provides new details in the velocity structure. These details are highly valuable for structural interpretation.

5.8. Uncertainties in the Optimal Blocky Velocity Model

[41] The quality of the final velocity model (optimal blocky model) is assessed by using the fit between predicted and observed traveltimes to estimate the error of velocity and interface node related to the ray coverage (Figure 18). According to Zelt [1999], there are two main sources of errors when working with coincident MCS and WA data: the first one is related to the mismatch between MCS reflectors and their corresponding WA reflection/refraction interfaces. To reduce this potential source of error, we correlated only the most prominent MCS reflectors with WA interfaces as explained below. A second source of errors could be associated with the inherent anisotropy of earth materials. However, during the combined inversion of MCS and WA data, we modify WA velocities to account for near-vertical incidence reflections. As Zelt [1999] indicated, the velocities in the isotropic model would correspond to a weighted average of fast and slow velocities, which depends on the subhorizontal and subvertical rays that sample each point of the velocity model. During RSTTI processing, we noticed that vertical velocities are $\sim 10\%$ lower than horizontal velocities, suggesting a weak anisotropy regime [Thomsen, 1986]. For this reason, the anisotropy of the materials may have an weak influence on the error analysis and is not taken into account in our processing sequence. To estimate the uncertainties of the velocities and interface depths and to asses whether the blocky velocity model inferred from the joint inversion of normal-incidence reflection and refraction and wide-angle reflection traveltimes provides improved model compared to that inferred from wide-angle traveltimes only, we followed the single parameter uncertainty test proposed by Zelt and Smith [1992]: The value of one model parameter in the optimal blocky velocity model is perturbed and held fixed. Then, traveltimes are inverted for all others parameters that could be dependent on the examined one. The size of the perturbation is increased until the reconstructed model is unable to fit the observed times. The maximum size of the perturbation that allows fitting the observed traveltimes is taken as the estimation of the absolute uncertainty for this parameter. This procedure is repeated for several representative model parameters. First, we applied this uncertainty analysis using only refraction and wide-angle reflection traveltimes (Figures 19a and 19b). Second, we repeat this analysis using both normal-incidence reflection and refraction and wideangle reflection traveltimes (Figures 19c and 19d). The nodes with velocity error <0.5 km s⁻¹ and depth error <0.5 km are considered well resolved and present low uncertainty [Zelt and Smith, 1992]. Comparison of Figures 19a and 19c show that velocity uncertainties in the subduction channel (SC) were decreased when the joint inversion of normalincidence reflection and refraction/wide-angle reflection traveltimes is performed. However, the velocity uncertainties have locally increased after the joint inversion, as, for example, at the top of the upper layer of the oceanic crust (OCA) near a 8–10 km depth (Figure 19c). Such worsening may be related to the locally poor WA ray coverage (Figure 18) and inaccurate picking on the normal incidence traveltimes on the stack section. We can also note that OWB presents a slight increase of the velocity uncertainties which



Figure 16



Figure 17. Modeling of the wide-angle reflection from the splay fault and the top of subducting plate. (a) Ray tracing for the WA reflection R2 from the splay fault and fit between corresponding observed (pink lines) and synthetic (black lines) times. (b) Ray tracing for WA reflection R3 from the top of the subducting plate and fit between observed (green lines) and synthetic (black lines) times.

may be related to uncertainties in the picking of horizons in a poorly reflective zone of the time stack section and the numerical limitations of the modeling [*Zelt and Smith*, 1992] to estimate errors in very complex media. A significant decrease of the uncertainty of the interface depths is obtained (Figures 19b and 19d) when the joint inversion is used (see the depth nodes defining the top of the layers SC, OWB, IWB1, IWB2).

6. Results and Discussion

[42] The optimal blocky velocity model, which consists of seven layers (see Table 3 and Figure 3), has been superimposed on the PSDM image (Figure 20) to help discriminating and discussing the main margin structures.

6.1. Sedimentary Cover

[43] The sedimentary cover (SD) is interpreted from its well-bedded seismic facies in MCS data and relatively low velocities ($<3 \text{ km s}^{-1}$). It divides into trench, margin slope and fore arc basin sediments. At the trench, sediment thickness ranges from 1.3 km at the seaward end of the section to 3.2 km near the deformation front (Figure 16, km 0). These thicknesses are within the 1–3 km range proposed by *Gailler et al.* [2007]. P wave velocity of trench sediment increases from 1.79 km s⁻¹ on average at the seaward limit of the line to 2.59 km s⁻¹ at the deformation front. This lateral velocity increase may be related to a change in sediment composition from pelagic-rich sediments away

from the trench to relatively more sandy and compacted sediment near the trench axis, in agreement with white chalk dredged from the summit of the eastern flank of the Yaquina graben, and sandy turbidites cored from the Esmeraldas deep-sea turbidite system in the trench axis during the Amadeus cruise [Collot et al., 2005]. Moreover, on the basis of the PSDM, Collot et al. [2008] show that the 3-kmthick trench fill is underlain by a 1- to 1.5-km-thick tectonic graben filled with 3.8-4.5 km s⁻¹ reflective sequences. Across the margin slope and fore-arc basin, the sediment thickness varies from less than 0.5 km beneath the Esmeraldas canyon to 2.5 and 3 km in the western and eastern depot centers of the basin, as indicated by Gailler et al. [2007]. Velocity of layer SD is around 1.79 km s⁻¹ along the layer upper part of the fore-arc basin, a value that is consistent with poorly consolidated muddy sediment cored in the basin [Collot et al., 2005]. Layer SD velocity increases to 2.9 km s⁻¹ at its base, and 2.5 km s⁻¹ on average over the outer margin wedge between km 0 and 18 along the section (Figure 16 and Table 3). The 0.39 s⁻ vertical velocity gradient of the layer (Table 3) supports downward consolidation of sediments due to increasing lithostatic pressure. The seaward velocity increase from 1.79 to 2.5 km s⁻¹ likely reflects consolidated sediment outcropping at the seafloor of the margin slope, a situation that is consistent with sediment compaction caused by compressive tectonic at the margin front (Figure 16, km 0-8), and shallow sediment stripping due to gravity sliding on the margin slope (Figure 16, km 10-18) [Ratzov et al., 2007].

6.2. Inner Wedge Basement

[44] The inner wedge basement comprises upper (IWB1) and lower (IWB2) crustal layers. The IWB1 is a 4- to 7-kmthick layer characterized by a 4.52 km s⁻¹ average velocity at the top of the layer, and 6.03 km s^{-1} near its base (Table 3). During the RSTTI modeling procedure, a floating reflector was needed. This reflector is consistent with phase Rf, which is locally identified on OBSs 120–128 (e.g., Figures 6–8). Reflector Rf became the lower boundary of the IWB1. This boundary dips slightly landward and shows a 3-km-high offset at km 60 (Figure 3), suggesting a steeply dipping crustal fault. The fault, which is interpreted as a strike-slip fault zone on the basis of the detailed PSDM image [Collot et al., 2008], puts the 7-km-thick seaward segment of layer IWB1 into contact with its only 4-km-thick landward counterpart. WA resolution does not allow identifying a velocity discontinuity at the boundary between layers IWB1 and IWB2, despite the wide-angle reflection Rf. Ray tracing shows that refracted arrival Pg2 (Table 1) illuminates only the shallower (\sim 3 km) part of IWB2, deeper velocities being constrained by reflections R2 and PmP and up going refracted Pc arrivals. IWB2 presents a 6.03 km s⁻¹ average velocity at the top of the layer and 6.60 km s⁻¹ at its base

Figure 16. Normal incidence traveltime modeling/inversion. (a) Stack time section. (b) Stack time section with picking of the major reflectors To, SF, B, and IB. (c) Example of normal incidence ray tracing from the interfaces of the optimal blocky velocity model. (d) Comparison between the geometry of the reflectors before and after the joint inversion of normal-incidence reflection and refraction/wide-angle reflection traveltimes. Note major shallowing of the downgoing oceanic crust in the deep part of the model (red curves). (e) Fit between observed normal-incidence reflection traveltimes (blue lines) and those computed in the optimal blocky velocity model of Figure 3. Box delineates the region where major reflectors where picked on the stack time section (Figure 16b). TWT, two-way traveltime.



Figure 18. Ray coverage for the optimal blocky model. The gray scale represents the number of rays per cell. Each cell is a rectangle of dimensions x = 0.5 km and z = 0.25 km. Note a good coverage near the base of outer wedge basement and a low coverage in oceanic crust (for example, the white zone below OBS 116 and at the depth z = 10-12 km).

near a 15-km depth, 60 km from the trench (Table 3). IWB1 and IWB2 velocities taken collectively corroborate the oceanic nature of the margin basement in the studied area as proposed by *Gailler et al.* [2007]. Our new analysis supports an oceanic plateau origin for the margin basement along line SIS-44. This line cuts across the Gorgona terrane [*Cediel et al.*, 2003] that outcrops on Gorgona Island (Figure 1) and consists of a suite of upper Cretaceous oceanic rocks and serpentinized peridotites that are part of an accreted oceanic plateau [*Dietrich et al.*, 1981; *Storey et al.*, 1991; *Kerr et al.*, 1996]. The total thickness of layers

IWB1 and IWB2 amount to a minimum of 14 km making the margin structure incompatible with normal oceanic crust, unless original oceanic crust was stacked during a compressive event. Conversely, the high- (0.24 s^{-1}) and low- (0.08 s^{-1}) velocity gradients within IWB1 and IWB2, respectively, (Table 3) show close resemblance with those of the upper and lower crusts of oceanic plateaus [*Charvis et al.*, 1995]. Considering the segment of layer IWB1 located between km 40 and 60 as representative of the layer, its thickness (5–7 km) and average velocity (5.28 km s⁻¹) compare relatively well with characteristics of the upper



Figure 19. (a) Velocity error and (b) geometry error at model nodes for the model constrained by WA data only. (c) Velocity error and (d) geometry error at model nodes for the optimal blocky velocity model constrained by WA and MCS data. See text for explanations.



Figure 20. Optimal blocky velocity model with isovelocity contours in km s⁻¹ superimposed on the optimal PSDM seismic image. Basement acronyms are defined in Table 3. SF, splay fault; FTS, frontal tectonic sliver; Esm C, Esmeraldas canyon; LZ, zone of landslide; Gr, summit graben.

crust of the Iceland oceanic plateau [Brandsdóttir et al., 1997; Staples et al., 1997; Dardyshire et al., 1998], and Kerguelen-Heard Plateau [Charvis et al., 1995], which both show 5- to 8-km-thick upper oceanic crust with 5.2-5.5 km s⁻¹ average velocities. The thickness and velocity structure of layer IWB2 cannot be directly compared to the lower crust of an oceanic plateau because a significant part of the foundation of IWB2 was likely removed by subduction erosion [Collot et al., 2008], only leaving a wedge of rock that samples the very upper section of the original plateau lower crust. Nevertheless, the velocity structure of the first 5-6 km of layer IWB2 that shows average velocities of 6.32 km s⁻¹, compares well with average velocities of 6.5–6.7 km s⁻¹ at the top part of the lower crust of the Iceland and Kerguelen-Heard oceanic plateaus. We conclude that the margin basement imaged in line SIS-44 is compatible with the remains of an oceanic plateau.

6.3. Outer Wedge Basement and the Splay Fault

[45] MCS data shows low-amplitude reflections from the outer wedge basement, suggesting weak impedance contrast at this interface, and indicating that seismic waves had little penetration in this structure. Thus, velocity and thickness

information depends almost exclusively on WA data. P wave velocity is constrained by refracted phase P2 and up going refracted phase Pc (Table 3). Reflectors, associated with high-velocity contrasts delineate the outer wedge basement (OWB) (Figure 3). WA tomography shows OWB to be associated with a large-scale low-velocity zone [Gailler et al., 2007]. Our analysis supports, however, the possibility of two shadow zones (SFSZ and SCSZ in Figures 6 and 7) likely related to distinct low-velocity zones: a splay fault low-velocity zone and a subduction channel low-velocity zone (Figure 20). Therefore, the velocity model obtained with our procedure (Figure 20) shows OWB to have a more complex internal velocity structure than that proposed by Gailler et al. [2007]. OWB presents a maximum average thickness of 4.5 km (Table 3), shows a deep, high-velocity core $(5.0-5.5 \text{ km s}^{-1})$ (Figure 20), and is separated from the inner wedge basement by the splay fault interface (SF). The splay fault dips landward and soles out on the plate interface between 12 and 15 km depth as suggested by Gailler et al. [2007]. The SF is, however, closely associated with the low-velocity, wedge-shaped zone $(4.0-5.0 \text{ km s}^{-1})$ that dips landward, and is squeezed between the inner wedge basement and the

Table 3. One-Dimensional Parameterization of the Margin, Based on the Optimal Blocky Model^a

Model Unit	Upper Velocity (km s ⁻¹)	Lower Velocity (km s ⁻¹)	Average Velocity (km s ⁻¹)	Velocity Gradient (s^{-1})	Average Thickness (km)
Sedimentary cover (SD)	1.79	2.59	2.19	0.39	2.03
Upper inner wedge basement (IWB1) ^b	4.52	6.03	5.28	0.24	6.38
Lower inner wedge basement (IWB2) ^b	6.03	6.60	6.32	0.08	7.27
Outer wedge basement (OWB) ^c	3.72	5.00	4.36	0.29	4.44
Subduction channel (SC)	3.81	3.77	3.79	-0.04	0.93
Ocean crust layer A (OCA)	4.92	6.49	5.70	0.72	2.20
Ocean crust laver B (OCB)	6.50	6.85	6.67	0.09	4.00

^aUpper velocity is velocity in the upper part of the layer. Lower velocity is the velocity in the lower part of the layer.

^bOwing to ray covering and geometry, average velocity and thickness for IWB1 and IWB2 units are taken between x = 50 and 60 km.

^cOwing to the high variability, velocity and thickness for OWB unit are taken at x = 25 km.

outer wedge high-velocity core. The nature of the OWB remains unclear. It might be interpreted as an accretionary wedge on the basis of its wedge shape, relative position to IWB. This wedge should be older than the oldest overlying fore-arc basin sediment interpreted to be Eocene in age [*Marcaillou and Collot*, 2008]. However, OWB internal structure is not typical of an imbricated accretionary wedge. Alternatively, its high-velocity core suggests that OWB comprises oceanic rocks, which may belong to a different petrologic domain of an oceanic plateau than those of IWB1. In this hypothesis, the low-velocity wedge-shaped zone associated with the SF could result from rock alteration in response to fluids circulation and tectonic shearing along the splay fault. Fluids are substantiated by high-amplitude splay fault reflectivity of negative polarity [*Collot et al.*, 2008].

6.4. Subduction Channel

[46] Between km 24-44 and 6-8 two-way traveltimes, the MCS time section (Figure 2) revealed two discontinuous but strong reflectors bounding the subduction channel, and interpreted as the interplate décollement (De) and the top of the oceanic crust (To) [Collot et al., 2004]. However, these reflectors cannot be ascertained between km 14 and 24 along the MCS time section. Modeling the Subduction Channel shadow zone (SCSZ in Figures 5-7 and SZ in Figures 8 and 9) observed in the WA data supports a lowvelocity subduction channel located between the outer wedge basement and the underlying oceanic crust. A 3.7 to 4.0 km s⁻¹ velocity in the SC accounts for modeling upgoing refraction Pc and reflection PmP, with a variable thickness ranging from 1.3 km to less than the WA seismic vertical resolution (~ 0.9 km). Although SC velocities are necessarily lower than those of the overlaying basement rocks to fit WA data, the model is non unique because the SC velocity could not be univocally determined. The SC velocities would poorly characterize compacted underthrusting sediment at shallow levels of the SC, and underplated consolidated sediments possibly mixed with oceanic crust and margin basement fragments typical of subduction channels, at deeper levels. Evidence for reverse polarity reflections on the roof thrust and inside the SC supports fluid circulation at depths as deep as 10-12 km [Collot et al., 2008].

6.5. Oceanic Crust and Mantle

[47] The thickness of the mafic Nazca oceanic crust, which is constrained by the reflections R2 at the interplate and PmP at the Moho, is about constant at 6.2 km. This result refines the 6- to 7-km-thick oceanic crust obtained by Gailler et al. [2007] along the same seismic line. The crust divides into a lower-velocity, high-gradient upper layer OCA and a higher-velocity, low-gradient, lower layer OCB (Table 3). No velocity discontinuity marks the boundary between the two layers, which are separated by a change of velocity gradient. The velocity characteristics of the Nazca oceanic crust are compatible with typical velocity and thickness values compiled for a normal young oceanic crust by White et al. [1992], and with the ~ 20 Ma age assigned to this segment of the Nazca plate by Hardy [1991]. It is interesting to note that the OCA velocity vertical gradient and average velocity (5 km s⁻¹) remain constant from the western end of the section, up to km 34 along the line, at a 12 km depth, where OCA velocity

increases to 6.0 km s⁻¹. This lateral velocity increase with depth is not an artifact related to modeling the overlaying SC low-velocity zone, because the velocity increase was already pointed out on the tomographic model by Gailler et al. [2007]. A consequence might be a clue for incipient physical transformations of crustal oceanic rocks with depth. A 7.8 km s⁻¹ mantle velocity is constrained by Pn arrivals. The velocity contrast of ~ 0.8 km s⁻¹ across the Moho can account for the high amplitude of PmP reflections. No upper mantle velocity gradient was detected, possibly because Pn refractions do not penetrate deep enough into the mantle. The upper mantle velocity of 7.8 km s⁻¹ in the vicinity of the trench axis might indicate a small (\sim 12%) degree of serpentinization of ultramafic rocks as found offshore Nicaragua by Ivandic et al. [2008]. This process would result from widespread mantle hydration due to seawater circulating along bending-related normal faults of the subducting plate [Ranero et al., 2003].

6.6. Origin of the Margin Basement Rocks

[48] Our geophysical transect extends across the interpreted Gorgona oceanic terrane that outcrops on Gorgona Island, and forms the basement of the margin, offshore SW Colombia [Cediel et al., 2003]. Geochemical analysis of geological samples from Gorgona Island [Storey et al., 1991] supports an oceanic plateau origin for the Gorgona terrane. On the basis of seismic data, we show that the velocity structure of the margin basement is compatible with such an origin. The total thickness of layers IWB1 and IWB2 amount to a minimum of 14 km making the margin structure incompatible with normal oceanic crust, unless original oceanic crust was stacked during a compressive event. Conversely, the high- (0.24 s^{-1}) and low- (0.08 s^{-1}) velocity gradients within IWB1 and IWB2, respectively (Table 3), show close resemblance with those of the upper and lower crusts of oceanic plateaus. Considering the segment of layer IWB1 located between km 40 and 60 as representative of the layer, its thickness (5-7 km) and average velocity (5.28 km s^{-1}) compare relatively well with characteristics of the upper crust of the Iceland oceanic plateau [Brandsdóttir et al., 1997; Staples et al., 1997; Dardyshire et al., 1998], and Kerguelen-Heard Plateau [Charvis et al., 1995], which both show 5- to 8-kmthick upper oceanic crust with 5.2-5.5 km s⁻¹ average velocities. The thickness and velocity structure of layer IWB2 cannot be directly compared to the lower crust of an oceanic plateau because a significant part of the foundation of IWB2 was likely removed by subduction, only leaving a wedge of rock that samples the very upper section of the original lower crust. Nevertheless, the velocity structure of the first 5-6 km of layer IWB2 that shows average velocities of 6.32 km s^{-1} compare well with average velocities of 6.5-6.7 km s⁻¹ at the top part of the lower crust of the Iceland and Kerguelen-Heard oceanic plateaus. We conclude that the margin basement imaged in line SIS-44 likely is compatible with the remains of an oceanic plateau.

7. Conclusions

[49] We propose here a strategy that combines MCS data for shallow velocity estimation with WA data for deep velocity estimation to improve prestack depth migration (down to 20 km in depth) and build a well-constrained blocky velocity model. The construction of the blocky model includes an iterative correction of geometry and velocity of deep reflectors, using wide-angle data and integrating normal incidence times, so that a velocity model that explains all four kinds of arrivals (normal incidence, short-offset reflections and wide-angle reflections/refractions) is obtained. The resulting PSDM image together with the optimal blocky velocity model reveals the following:

[50] 1. A ~6.2-km thick Nazca plate oceanic crust with downdip velocity increase suggests incipient physical transformation of mafic rocks with depth [*Gailler et al.*, 2007]. A 7.8 km s⁻¹ upper mantle velocity supports some degree of ultramafic rock serpentinization in the bending region of downgoing plate.

[51] 2. Fine-scale sediment lateral velocity variations both across the trench and frontal margin slope account for lithologic variations, tectonic compaction, and mass wasting processes.

[52] 3. A two-layer velocity structure of the inner wedge margin basement is cut by a major subvertical fault system and is compatible with the upper and lower crusts of an oceanic plateau.

[53] 4. The outer wedge basement, which has significantly lower velocities $(4.0-5.5 \text{ km s}^{-1})$ than the inner wedge basement $(4.0-6.6 \text{ km s}^{-1})$ [*Gailler et al.*, 2007], shows a complex velocity structure dominated by a deep, highvelocity $(5.0-5.5 \text{ km s}^{-1})$ core, and a low-velocity zone $(3.8-5.0 \text{ km s}^{-1})$ associated with a major landward dipping fault that splays away from the plate interface. The lowvelocity splay fault zone may result from tectonic shearing and fluids migration along the fault.

[54] 5. A \sim 1.3-km-thick, low-velocity (3.5–4.0 km s⁻¹) subduction channel occurs beneath the outer wedge basement.

[55] 6. Physical properties of the subduction channel as well as its interaction with the splay fault may play an important role on basal erosion and underplating, as well as on the fault mechanics during the earthquake cycle as discussed by *Collot et al.* [2008] for the same transect.

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