Suppression of free-surface effects from multicomponent sea-floor data

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ABSTRACT

We use multicomponent sea-floor data to suppress receiver-side ghosts and free-surface multiples and to datum the results of ghosts and multiple suppression from the sea-floor to the free-surface. Both steps – deghosting and multiple suppression - are applied trace by trace to the *tau-p* transformed common receiver gathers. Datuming for 2D structures with inline lateral variations is performed in the same *tau-p* domain. Therefore effective suppression of aliasing in the *tau-p* transform is of crucial importance. In addition to the dip filter our antialiasing protection is based on simple scaling of the results of the transform for high frequencies. The synthetic and real data examples illustrate that the total scheme performs well and is computationally efficient.

INTRODUCTION

Our scheme for ghost and multiple suppression is essentially a combination of the work of Soubaras (1996) and a modification of the approach of Amundsen (1999). It is well known that the receiver-side ghost effects can be suppressed by weighted summation of pressure and vertical velocity records. The problem is to estimate the weighting filter. The idea of Soubaras is the following: The direct wave contribution S can be obtained as a linear combination of hydrophone and vertical geophone records. If the records are properly weighted, the result of this linear combination should be equal to zero (close to zero for real records) for times larger than the source-receiver travel-time plus the signature length. After defining the weighting filter, we decompose the hydrophone and vertical geophone records into the contributions of upgoing and downgoing P waves. The results of decomposition can be used directly to calculate the subsurface reflection response without any free-surface effects (Amundsen, 1999). For a 'locally' 1D medium, this response is calculated as RS', where S' is the spectrum of any chosen source signature including the phase shift due to the direct wave source-receiver propagation, while R is the ratio of upgoing and downgoing waves (in the frequency-slowness or frequency-wavenumber domain). In contrast to Amundsen, we use the estimate of R not for direct evaluation of primaries, but for prediction of multiples, which we then adaptively subtract from hydrophone and geophone records.

SUPPRESSION OF FREE-SURFACE GHOSTS AND MULTIPLES

The hydrophone and vertical-component geophone records can be decomposed into upgoing and downgoing waves after defining the calibration filter (Soubaras, 1996; Lokshtanov, 2000). The results of decomposition can be used to suppress all free-surface multiples (Amundsen, 1999). Below we rederive the result of Amundsen both for 1D and 2D structures using the invariant imbedding approach (Kennett, 1983), which gives straightforward physical insight into the operations applied. All expressions below are for plane harmonic waves. Therefore the first processing step should be the decomposition of the recorded traces into plane wave contributions. For a 'locally' 1D structure the total upgoing wavefield U has the following form, Figure 1:

$$U = \{R + Rr_f R + Rr_f Rr_f R + ...\}S = \{1 - Rr_f\}^{-1} RS \,, \tag{1}$$

where S is the source incident wavefield (with the source-side ghost), r_f is the reflection coefficient from the free-surface, while R is the generalized reflection coefficient from

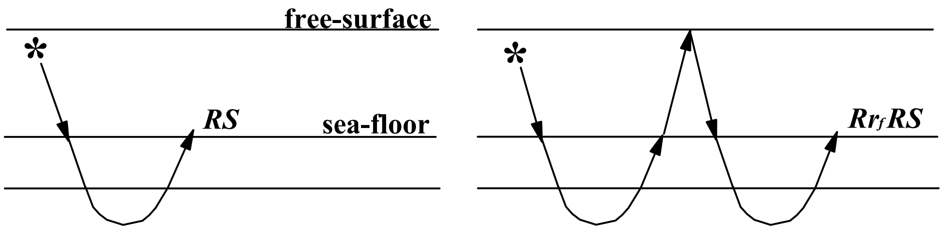

FIG. 1. 'Rays' for the first terms of (1).

below the sea-floor without any free-surface effects. Note that in the definition of both r_f and R the incident and reflected plane waves are 'measured' just above the sea-floor. Similarly, the model of the total downgoing wavefield D is:

$$D = \{1 + r_f R + r_f Rr_f R + ...\}S = \{1 - r_f R\}^{-1} S \tag{2}$$

From (1)-(2) we obtain the primary response RS' due to any chosen source signature S':

$$RS' = \frac{U}{D} S' \approx \frac{UD^*}{(DD^* + \varepsilon^2)} S' , \tag{3}$$

where ε is a regularization parameter. Essentially, formula (3) defines the scheme of Amundsen (1999). The great advantage of his scheme is that it does not require either source signature measurements, or optimization with respect to the signature parameters. At the same time the scheme does not account for misfit between the real data and the data model. We have modified his scheme in the following. We use the estimate of R, $R \approx UD^* / (DD^* + \varepsilon^2)$, not for direct evaluation of primaries, but for prediction of multiples. From (3) it follows that all free-surface multiples M in U are :

$$M = Rr_f U \text{, where } R \approx UD^* / (DD^* + \varepsilon^2) \text{ and } r_f = -\exp\{2i\omega qh\} \, . \tag{4}$$

In (4) h is water-layer thickness, while q is vertical slowness. After prediction of multiples M, we adaptively subtract them from U. The adaptive subtraction is performed trace by trace in the *tau-p* domain. A synthetic example is shown in Figure 2: input hydrophone data with all ghosts and multiples (left), after suppression of receiver-side ghosts (centre) and after suppression of multiples (right). The real data example is shown in Figure 3.

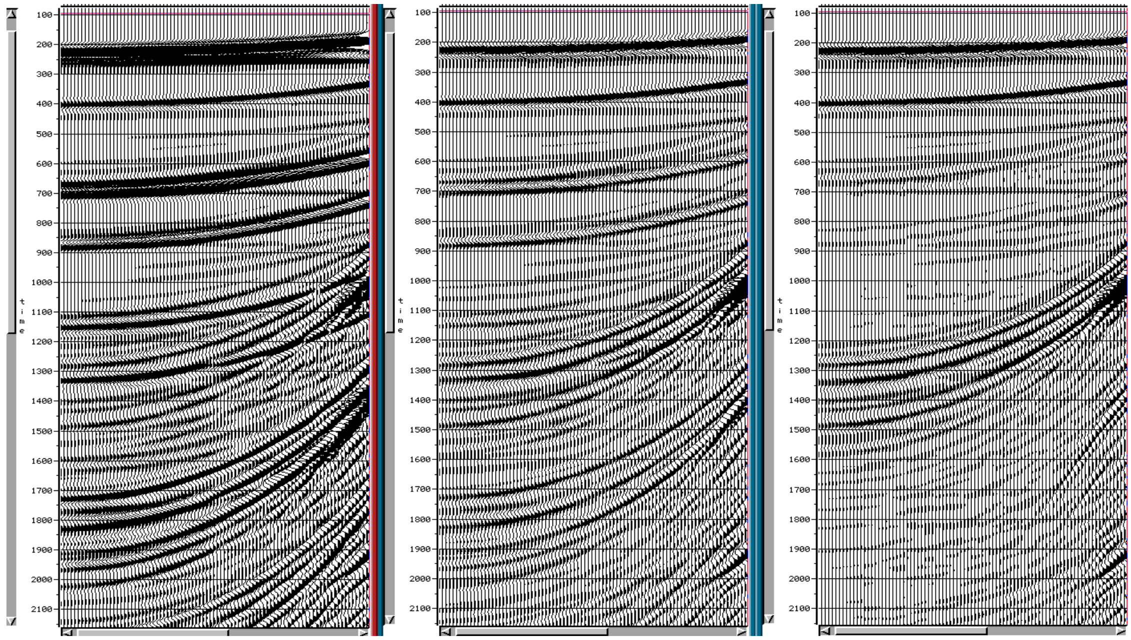

FIG. 2. Input data (right), after HZ merge (centre), after suppression of multiples (right)

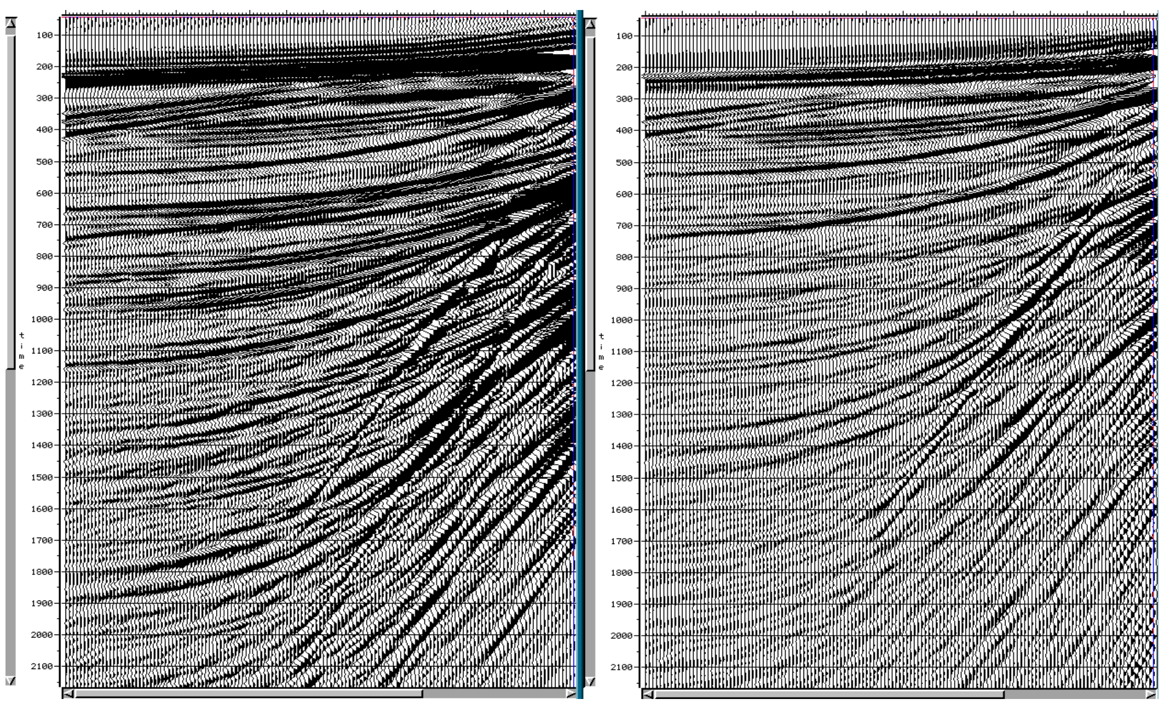

FIG. 3. Input hydrophone data (right) and after suppression of ghosts and multiples (left)

In a similar way we predict and subtract all free-surface multiples from the radial geophone components. Denote by TS the primary reflection response (with all internal multiples and conversions) for the radial component, where S is the direct source wavefield (with the source-side ghost; recall that S is defined just above the sea-floor), while T is the product of the geophone response and the medium response. T is the radial component data (frequency and slowness dependent) due to an incident wavefield of unit amplitude; note also that the radial component is registered just below the sea-floor. With this notation the model of the radial component X with all free-surface effects is:

$$X = T\{1 + r_f R + r_f R r_f R + \ldots\}S = T\{1 - r_f R\}^{-1} S \,. \tag{5}$$

From (5) we get the 'multiple part' of the radial component: $M_x = R r_f X$, where R is defined in (3). The synthetic example of multiple suppression from the radial geophone component is shown in Figure 4.

Finally, note that for 2D structures the data model (1), (2) and (5) is still valid if all terms in these formulas are matrices, where different rows and columns correspond to different slownesses from the receiver- and the source-side respectively. Similarly to (3), the subwater reflection matrix R without any free-surface effects is: $R = UD^{-1}$.

DATUMING OF *TAU-P* GATHERS FOR 2D STRUCTURES

After multiple suppression we can datum sources from the free-surface to the sea-floor or recordings from the sea-floor to the free-surface. Datuming of sources is trivial and is defined by the phase shift $\exp\{-i\omega qh\}$ for each p-trace of *tau-p* transformed common-receiver gathers. The disadvantage of this procedure is that it corrupts the reflections from the very shallow part of the structure. Datuming of receivers from the sea-floor to the free-surface is not as trivial, simply because the p-parameter of the Radon transform of common-receiver (CR) gathers is a source-side ray parameter, which is generally not equal to the receiver-side ray parameter. In the following we will assume locally 1D water-bottom and arbitrary 2D structure below it. The input Radon transformed CR gathers $D(p_s, x)$ for receiver position x can be represented as follows:

$$D(p_s, x) = \frac{\omega}{2\pi}\int R(p_s, p_r)\exp\{i\omega(p_r - p_s)x\}dp_r\,, \tag{6}$$

where $R(p_s, p_r)$ is the complex (frequency dependent) amplitude of the reflected plane wave with slowness p_r due to the incident plane wave with slowness p_s. With these notations the datumed Radon transformed CR gathers $D_g(p_s, y)$ for receiver positions y can be obtained as:

$$D_g(p_s, y) = \frac{\omega}{2\pi}\int R(p_s, p_r)\exp\{i\omega[(p_r - p_s)y + q_r h]\}dp_r = \\ \frac{\omega}{2\pi}\int D(p_s, x)\left\{\int \exp\{i\omega[(p_r - p_s)(y - x) + q_r h]\}dp_r\right\}dx, \tag{7}$$

where h is the 'local' water-bottom depth, while $q_r = (1/c^2 - p_r^2)^{1/2}$ is the vertical receiver-side slowness**;** c is water velocity. The inner integral in the curly brackets can be calculated by the stationary phase approximation. For each pair x, y the stationary point

p_r^{st} corresponds to a simple relation $x - y = h\,\mathrm{tg}\,\alpha_r$, where $p_r^{st} = c\sin\alpha_r$. α_r is the vertical angle for a ray between the receiver at point *x* on the sea floor and the receiver at point *y* at the free-surface. Figure 5 shows a stack of hydrophone data (left) and of results of ghost and multiple suppression (right). Both stacks are after *tau-p* datuming of the recordings from the sea-floor to the free-surface. A similar comparison with a stack from the vertical component data is given in Figure 6.

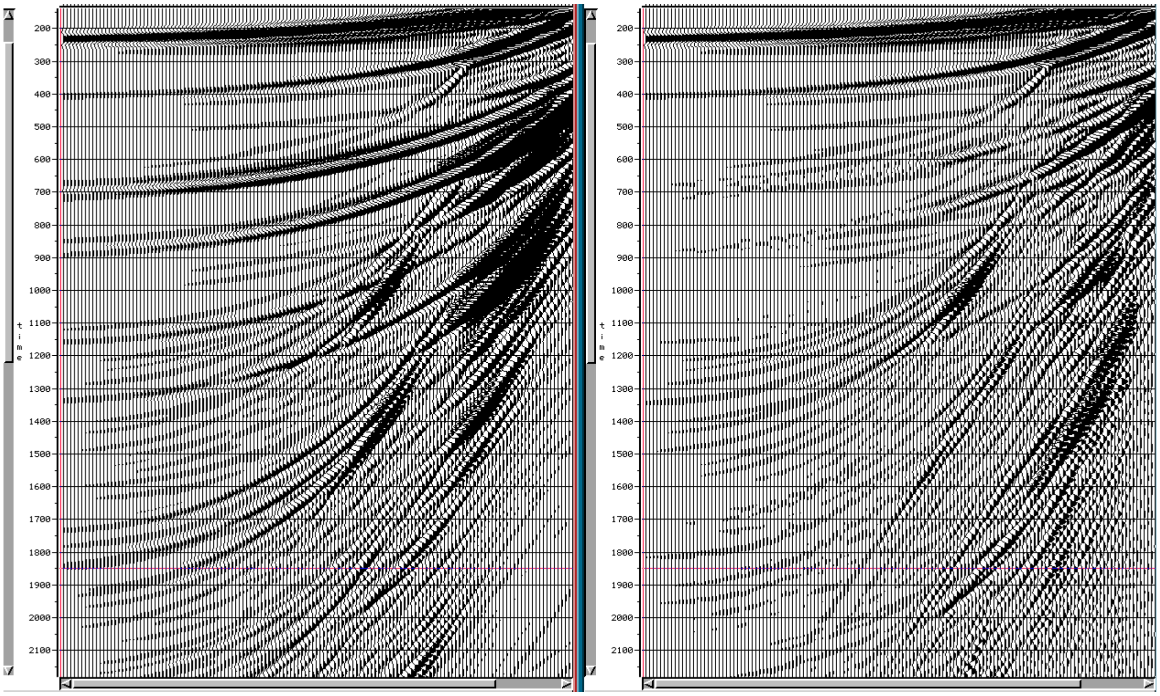

FIG. 4. Input radial component (left) and after suppression of ghosts and multiples (right)

ANTIALIASING IN THE RADON TRANSFORM

The Radon transform of CR gathers requires the use of both positive and negative dips for both positive and negative offsets. Negative dips for positive offsets (and vice versa) produce strong aliasing noise. The problem of aliasing in the Radon transform is relatively well understood (Turner, 1990) and can be summarized as follows. Denote by p_A and by p_R the actual dip (horizontal slowness) of an event present in the data and the dip used in the forward Radon transform. From the condition of constructive interference we get that for frequency *f* the combination p_A and p_R produces aliasing if $(p_A - p_R)\cdot\Delta x = nT$, where $T = 1/f$, Δx is the trace interval and *n* is an arbitrary integer. Suppose that we apply a conventional dip-dependent filter: $f_{\max}(p_R) = 1/2p_R\Delta x$. From the aliasing condition above it follows that such filtering does not prevent aliasing either for small or large slownesses in the transform. To avoid aliasing for small slownesses $(p_R \approx 0)$ we have to apply a low-pass filter to the data, so that $f_{\max} = 1/p_{\max}\Delta x$, where $p_{\max} = \max|p_A|$. For large slownesses the aliasing is still produced by a combination $f = f_{\max}(p_R)$ and $p_R = -p_A$. A hardly acceptable solution to the problem is further reduction of $f_{\max}(p_R)$ (for source spacing 25*m* and maximum dip in the data due

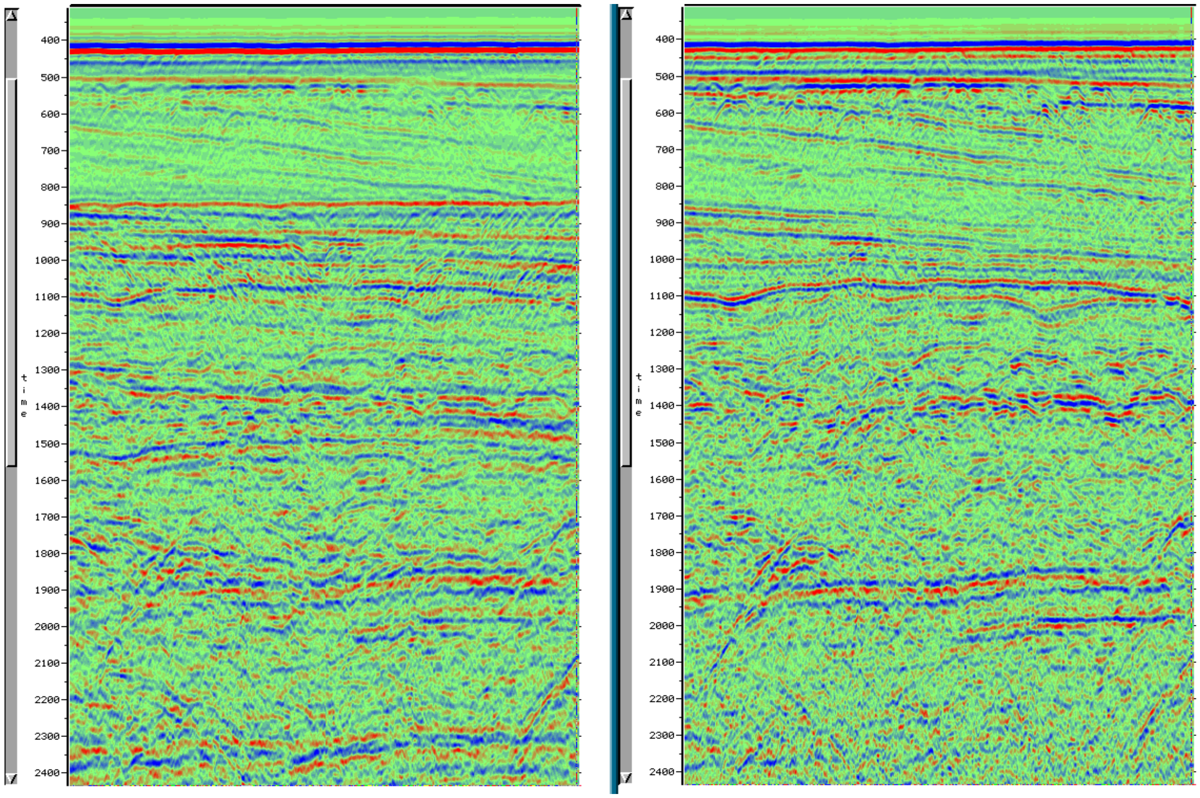

FIG. 5. Stack of hydrophone data (left) and of results of ghost and multiple suppression (right).Both stacks are after *tau-p* datuming of the recordings from the sea-floor to the free-surface.

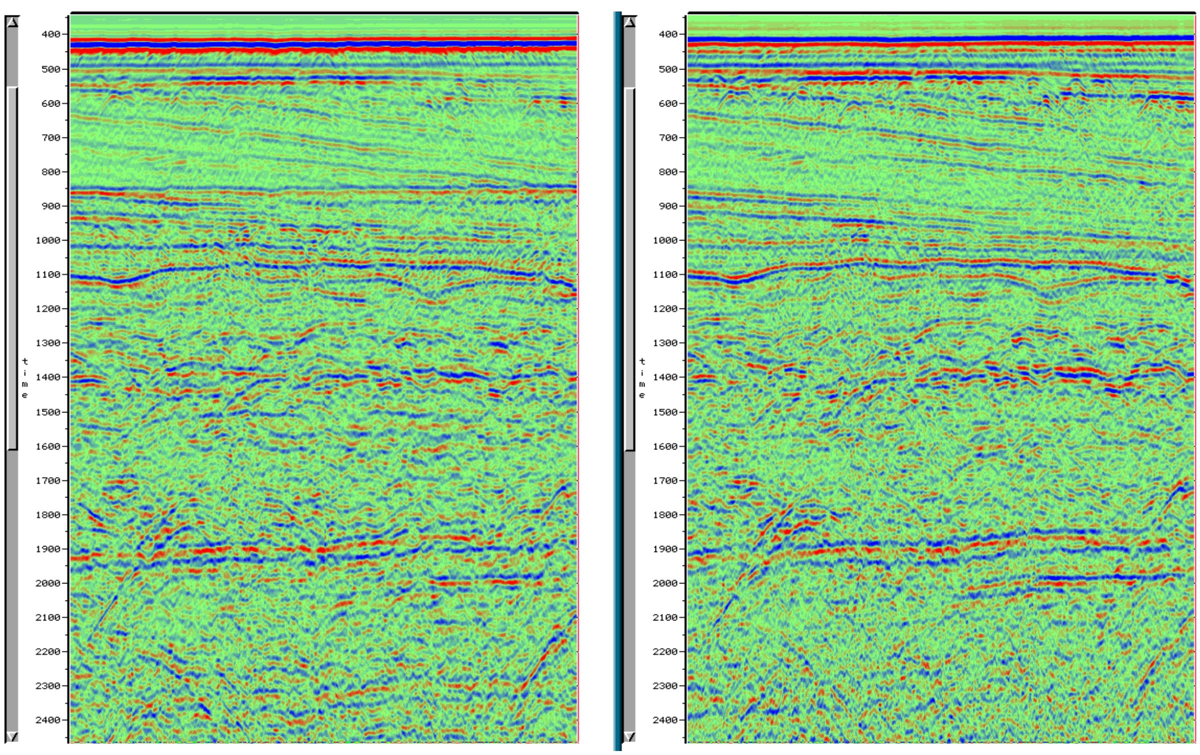

FIG. 6. Stack of vertical geophone component (left) and of results of ghost and multiple suppression (right).

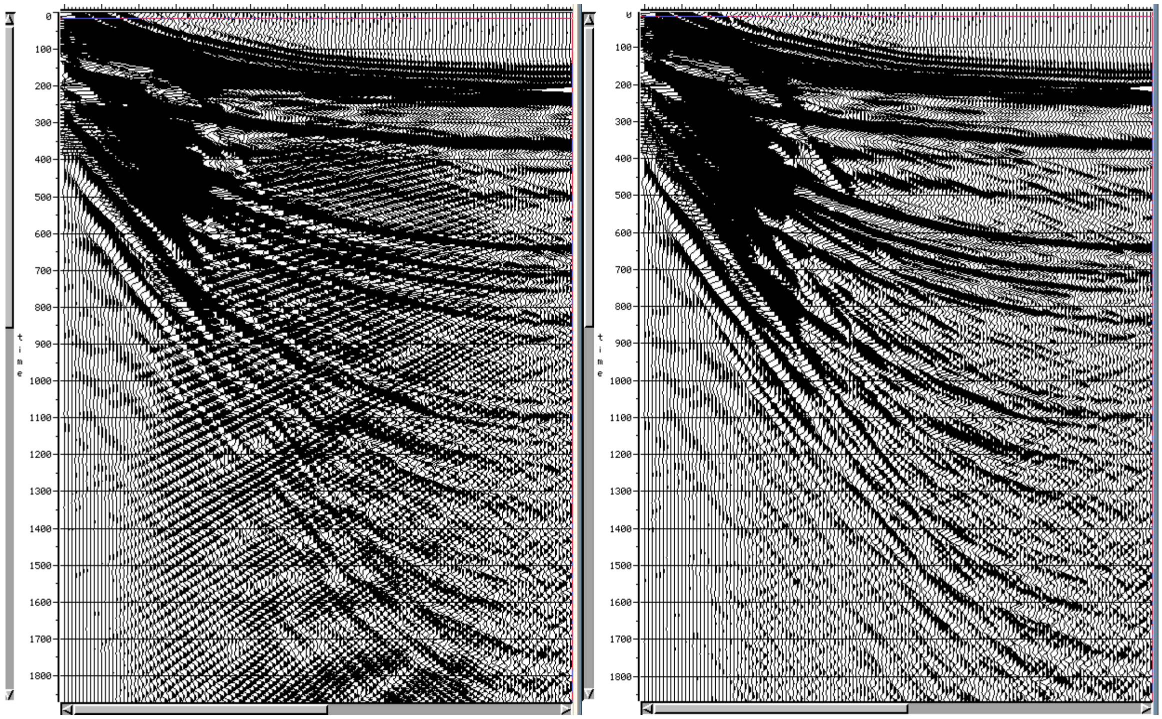

FIG. 7. Radon transform result of real CR gather with low-pass and dip-dependent filter (left) and with additional antialiasing protection (right).

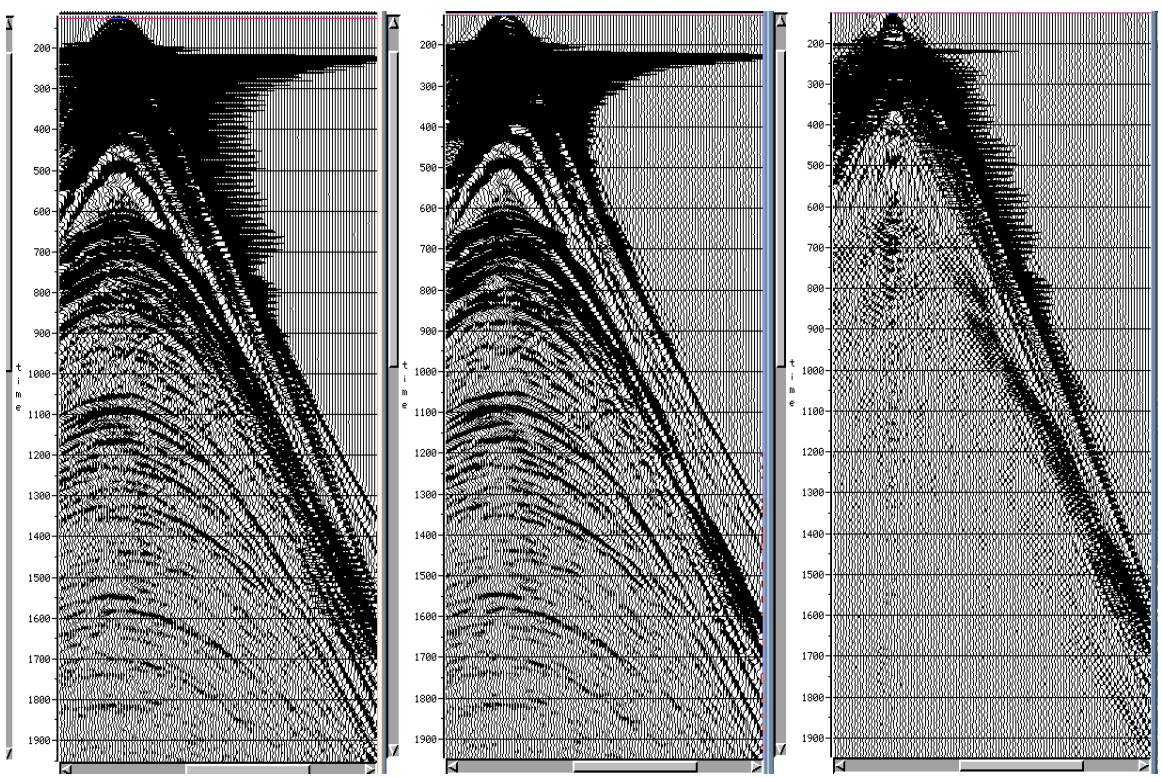

FIG. 8. Input CR gather (left), after forward and inverse Radon transform with additional antialiasing protection (centre) and the difference (right). The difference is first of all due to high frequency part of strongly aliased multiples.

to the sea-floor reflection, $f_{\max}(p_{\max}) \approx 30Hz$). Another alternative is to scale the transform results for high frequencies based on the results for low frequencies (see, for example, Herrmann et. al., 2000; Denisov and Finikov, 2001). Our simple scaling strategy is the following (Lokshtanov et al., 2002). First estimate the wavelet amplitude spectrum $S(f)$. Then using only positive offsets calculate the transform both for positive and negative dips. For each negative *p* trace calculate the transform amplitude spectrum $A(f)$ and evaluate the average ratio $Q_{low} = \int_{f_1}^{f_2} A(f)df \Big/ \int_{f_1}^{f_2} S(f)df$ in the low (nonaliased) frequency band (f_1, f_2). The ratio represents the AVO weighted number of traces tangent to the dip *p*. Finally, for each frequency in the 'high-frequency' range calculate $Q_{high} = A(f)/S(f)$. If Q_{high} is larger than Q_{low} scale the Radon transform results so that $Q_{high} = Q_{low}$. Do the same for negative offsets (with antialiasing protection for positive dips) and sum the Radon transform results from positive and negative offsets. The procedure is efficient, fast and accurate, Figures 7-8.

CONCLUSIONS

We use Radon transformed CR gathers to suppress the free-surface effects from multicomponent sea-floor data and to datum the results from the sea-floor to the free-surface. Efficient suppression of aliasing in the transform is of crucial importance. In addition to the dip filter, our antialiasing protection is based on simple scaling of the results of the transform for high frequencies. The synthetic and real data examples illustrate that the total scheme performs well and is computationally efficient.

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ACKNOWLEDGEMENTS

Thanks to Jan Petter Fjellanger (Norsk Hydro Research Centre, Bergen) and to Michael Denisov and Dmitri Finikov (GeotechSystem, Moscow) for stimulating discussions and technical assistance. Thanks to Norsk Hydro for permission to publish the paper.