trench line has occurred also during the 26/12 earthquake. For example, geodetic data recorded through global positioning system show a WSW to south-westwards smaller shift of trench line by about 4.8 m of the Andaman Islands, and a relatively larger shift by about 6.5 m towards south off the Nicobar region.

The western coastline of the Andaman–Nicobar Island is fairly straight and gentle in comparison to the notch and ophiolite-bearing eastern coastline. Tilting and gradual submergence of the island group westwards and consequent uplift of the Andaman Ridge since early Oligocene– Late Eocene, probably facilitated formation of the barrier reef along the west coast. Our 3D studies underway add to this understanding, showing dislocation and shifting of the Andaman subduction trench line towards west. This conforms to the earlier assessment for a general westward shift of the Andaman trench with every episode of subduction of the plate. Furthermore, the continuing aftershocks and recent volcanism suggest that the geodynamics in the arc areas of the Sumatra–Andaman belt is undergoing significant changes. It is time a holistic assessment of the entire Asian plate geometry is made.


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Physical parameters of hydrated sediments estimated from marine seismic reflection data

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Gas-hydrates in marine sediments can be identified on multi-channel seismic data by an anomalous bottom simulating reflector (BSR), often associated with the base of hydrate stability field. Physical parameters like porosity, density, thermal conductivity, temperature, geothermal gradient, hydrate saturation, electrical resistivity and heat flow provide useful inputs to understand several issues related to hydrates exploration. To determine these parameters by employing relevant techniques at depth below the sea water is not only difficult but also expensive. Here, we present case studies to derive these parameters with the help of BSR, identified on seismic sections from two completely different tectonic/geological areas of (i) the Makran (Arabian Sea) and (ii) the Cascadia (Pacific Ocean) margins. From the available velocity–depth model in the Cascadia margin, we determine the background velocity–depth function (without hydrates and free-gas), which is used to estimate the variation of density, porosity and hydrate saturation with depth successively. The average bulk density, seafloor porosity and maximum hydrate saturation in the Cascadia margin are calculated at 1.70 g/cc, 63.8% and 22% respectively. From the porosity–depth function and the gradient of hydrate saturation, we determine the variation of thermal conductivity and resistivity with depth. The average resistivity, thermal conductivity and heat flow in the Cascadia margin are determined as 1.13 Ω-W/m/K and 62.85 mW/m² respectively. Since the velocity model in the Makran accretionary prism across the BSR shows wide variation, we change the approach of deriving the above physical parameters. First, we determine the seafloor density from the seismic velocity and hence the seafloor porosity. Then the porosity–depth function is determined using Aby’s law and the compaction factor available for the sediments. Porosity is then converted into density using an empirical relation between porosity and density. The remaining procedure is the same as that used for the Cascadia margin. Seafloor porosity, average density, resistivity, thermal conductivity, heat flow and maximum hydrate saturation are calculated as 54%, 1.98 g/cc, 1.96 Ω-m, 1.268 W/m/K, 43.55 mW/m² and 13% respectively, for the Makran region. The estimated physical parameters in both the margins match well with the available results. We also estimate errors

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in physical parameters, assuming ±5% error in the available velocity model.

Keywords: Bottom simulating reflector, heat flow, hydrate saturation, porosity, resistivity, seismic velocity, thermal conductivity.

GAS-HYDRATES are ice-like crystalline solids of water and low-molecular weight hydrocarbons (mainly methane) that form at high pressure and low temperature. They are found in the sediments of permafrost within depths ranging from 130 to 2000 m and outer continental margins where water depth exceeds 500 m. The most commonly used marker for detecting gas-hydrates using seismic data is an anomalous reflector, known as the bottom simulating reflector (BSR), that shows high negative amplitudes and opposite polarity with respect to the seafloor reflections. Gas-hydrates have attracted the scientific community due to their widespread occurrences in nature, potential as future energy resources, possible role in climate change and submarine geohazards. To understand these aspects, determination of various physical parameters like porosity, density, hydrate saturation, thermal conductivity, temperature, resistivity and heat flow associated with the hydrate reservoirs provide essential inputs. However, measurement of these parameters using probes at depth below the sea water is not only difficult, but also expensive. Here we make an attempt to derive various parameters from the seismic velocity–depth function associated with the BSRs, identified in two distinctly different regions of the Cascadia margin in the Pacific Ocean (Figure 1) and the Makran accretionary prism in the Arabian Sea (Figure 2). Presence of gas-hydrates increases the seismic velocity that again depends on concentration and distribution, whereas free-gas reduces the seismic velocity considerably. Minshull et al. showed that the seismic velocity of gas-hydrated sediment increases with hydrate saturation. However, the P-wave velocity for free-gas saturated sediment decreases appreciably up to a few, say 5%, and above that the velocity remains almost unaffected by gas saturation. In other words, we cannot quantify gas saturation more than 5%, hence we restrict the approach of calculating various physical parameters up to the depth of the BSR only. First we determine the background or reference functions in the absence of hydrates and free-gas. All the equations used here are under the assumption of homogeneous and isotropic media.

Location of the multichannel seismic line L89-08 (near ODP site 889) in the northern Cascadia margin off Vancouver Island is shown in Figure 1 a. A strong BSR is observed in the reflection seismic section at 2.05 s TWT (Figure 1 b). Velocity as a function of TWT, derived using the waveform inversion at CDP 3100, is shown in Figure 1 c. The velocity–depth function (Figure 3 a) shows a monotonous increase in velocity from a depth of 125 mbsf and reaches the maximum value of 1.9 km/s at the BSR (215 mbsf). We determine various background parameters up to the depth of 350 mbsf. The velocity–depth function (Figure 3 a) shows that there exists a linear velocity trend (shown by hair line) if the effects of hydrates and free-gas are omitted. From this background velocity trend, P-wave velocities at the seafloor (lying at a water depth of 1233 m) and at 350 m mbsf are picked as 1.515 and 1.72 km/s respectively. The background velocity–depth function is represented as

\[ V_p(Z) = 1.515 + 0.59Z, \]

where \( Z \) is the depth below the seafloor and 0.59 km/s/km is the velocity gradient. If we consider ±5% error in velocity estimation, then the lower-bound and upper-bound background velocity–depth functions are expressed as

\[ V_p(Z) = 1.44 + 0.55Z, \]

\[ V_p(Z) = 1.59 + 0.62Z. \]

Using the background velocity–depth function given by eq. (1), we determine the background density–depth function (solid line in Figure 3 b) using Wang’s relation between the P-wave velocity \( V_p \) and the density \( \rho \) as

\[ V_p = c_p \phi^{b_p}, \]

where \( c_p = 0.4692 \) and \( d_p = 2.3229 \) are correlation constants for unconsolidated sediments. Marine sediments up to a few hundred metres below the seafloor where hydrates are stable, are unconsolidated in nature. The calculated densities at the seafloor and at 350 mbsf are 1.65 and 1.75 g/cc respectively, and hence the average bulk density of the sediment up to 350 mbsf is 1.70 g/cc. Dotted and the dashed lines (Figure 3 b) represent the lower and upper bounds of background density–depth functions calculated using eqs (2) and (3) respectively. The corresponding average densities are 1.66 and 1.74 g/cc, and seafloor densities are 1.62 and 1.69 g/cc respectively. The densities at 350 mbsf are 1.71 and 1.79 g/cc respectively.

The background velocity–depth function can be translated into the background porosity–depth function using any suitable velocity–porosity equation. To determine the porosity at the seafloor, we prepare a nomogram of P-wave velocity \( V_p \) vs sea floor porosity \( \phi \) as described below. For a given porosity \( \phi \), \( V_p \) can be calculated using the two-phase weighted equation of Nobes et al. as

\[ \frac{1}{V_p} = \frac{W \phi}{V_1} + \frac{(1-W \phi)}{V_2}, \]

where \( W \) is the weighting factor, taken as 1.2 due to the unconsolidated nature of marine sediments. \( V_1 \) and \( V_2 \) are calculated as follows:
where $V_m$ and $V_w$ are the $P$-wave velocities of sediment matrix and water respectively, $\rho_s$, $\rho_m$, and $\rho_w$ are densities of the sediment, matrix and water respectively. Here, $V_m = 4.37$ km/s, $V_w = 1.5$ km/s, $\rho_m = 2.68$ g/cc and $\rho_w = 1.05$ g/cc are used for clay-rich sediments in the area. The value of seafloor density ($\rho_s$) is already calculated as 1.65 g/cc. By varying the seafloor porosity between 20 to 80% at an interval of 1%, we calculate corresponding $V_p$ using eqs (5)–(7), shown as a solid line (nomogram) in Figure 4a. Corresponding to the seafloor velocity of 1.515 km/s (Figure 3a), seafloor porosity is determined as 63.8%. Similarly, we prepare nomograms of $V_p$ vs seafloor porosity corresponding to the lower (1.62 g/cc) and upper (1.69 g/cc) bounds of densities (Figure 4a). The lower and upper bounds of seafloor porosities are calculated as 59 and 69.5% respectively.

To calculate the porosity at 350 mbsf, we prepare another nomogram of $V_p$ vs porosity (solid line in Figure 4b) using eqs (5)–(7) in a similar way with porosity varying between 20 to 80% at an interval of 1% for the density of 1.75 g/cc at 350 mbsf. Corresponding to the background velocity of 1.72 km/s at 350 mbsf, we calculate the porosity as 54%. Considering ±5% error in velocity estimation at 350 mbsf, we also prepare nomograms of $V_p$ vs porosity using the lower and upper bounds of densities.
Figure 3. a, Full waveform inversion result showing velocity variation with depth$^{5,6}$ in the Cascadia margin. The background velocity–depth function is shown by hair line. b, Background density–depth function (solid line) derived from the background velocity–depth function. Dotted and dashed lines represent lower and upper bounds of density–depth functions.

Figure 4. Nomogram of (a) velocity vs seafloor porosity for the Cascadia margin and (b) velocity vs porosity for the depth of 350 mbsf. c, Background porosity–depth function along with upper and lower bounds.
(Figure 4b). The lower and upper bound porosities are determined as 49.5 and 59% respectively.

Background porosity for nonconsolidated sediment varies nonlinearly with depth, and the value, \( \phi(z) \) at a given depth (Z) can be calculated using the Athy’s law\(^{13}\):

\[
\phi(z) = \phi_0 e^{-\left(Z/\lambda\right)}, \tag{8}
\]

where \( \phi_0 \) is the porosity at the seafloor and \( \lambda \) is the compaction factor.

Since we have calculated porosities at the seafloor and at 350 mbsf, we can determine the value of \( \lambda \) as 2.09 per km using eq. (8). Compaction factors for the lower and upper bounds are 1.99 and 2.14 per km respectively. The estimated porosity vs depth curve from eq. (8) (calculated at every 5 m) and the error bounds are shown in Figure 4c, from which we get the porosity as 57.6% at BSR (215 mbsf) for the Cascadia margin. The lower and upper bounds of porosity at BSR are calculated as 53 and 62.9% respectively.

Hydrates may be considered as a part of the pore fluid situated away from the grain contact, defined as the non-contact model or treated as a part of the matrix, defined as the contact model\(^{12}\). To explain the velocity variation of hydrated sediment, the three-phase weighted equation\(^{13}\) for non-contact model and the effective medium theory\(^{14}\) for contact model are used. As the study\(^{15}\) of elastic properties for hydrate-bearing sediments shows that the weighted equation better predicts the seismic velocities than the effective medium theory, we consider the concept of non-contact model. To calculate the hydrate saturation \((S)\) at the BSR or the maximum saturation, we prepare a nomogram of \( V_P \) vs \( S \) by varying the hydrate saturation from 0 to 80% at an interval of 1% (Figure 5) using the three-phase weighted equation\(^{13}\) as follows:

\[
\frac{1}{V_P} = \frac{W\phi(1-S)^N}{V_1} + \frac{1-W\phi(1-S)^N}{V_2}, \tag{9}
\]

where \( N \) is a constant that simulates the rate of lithification with the hydrate saturation. \( N = 1 \) is used here. \( V_1 \) and \( V_2 \) are calculated as follows:

\[
\frac{1}{\rho_n V_1} = \frac{1-\phi}{\rho_n V_n} + \frac{S\phi}{\rho_n V_n} + \frac{(1-S)\phi}{\rho_n V_n}, \tag{10}
\]

and

\[
\frac{1}{V_2} = \frac{S\phi}{\rho_n V_n} + \frac{(1-S)\phi}{V_w}. \tag{11}
\]

Here, \( \rho_n (= 0.92 \text{ g/cc}) \) and \( V_n (= 3.3 \text{ km/s}) \) are the density and P-wave velocity of pure hydrates respectively. \( V_m = 4.37 \text{ km/s} \) and \( \phi = 57.6\% \) are used. Corresponding to the maximum P-wave velocity of 1.9 km/s at the BSR, the hydrate saturation is calculated as 22%. The value matches well with the hydrate saturation (20–30%) reported earlier\(^{9}\) for the region. Assuming 0% hydrate saturation at about 125 mbsf and maximum saturation at the BSR, the saturation gradient of hydrates is calculated as 0.24% per m, the value of which is required to calculate parameters like thermal conductivity and resistivity of hydrated sediments. The lower and upper bounds of maximum hydrate saturation at the BSR are 21.3 and 22.5% respectively (Figure 5).

From the seafloor temperature vs water depth profile (Figure 6a) obtained from the depth of ocean-bottom and temperature data\(^{16}\), we calculate the seafloor temperature as 2.8°C corresponding to a water depth of 1233 m in the study area. From the pressure-temperature phase diagram\(^{6}\) (Figure 6b), we determine the BSR temperature as 16°C corresponding to the equivalent pressure (16.27 MPa) of water and sediment column up to the BSR. The difference in temperatures between the BSR and the seafloor divided by the BSR depth with respect to the seafloor produces the geothermal gradient of 61.4°C/km. The ±5% error in sediment-velocity causes little change in pressure from 16.19 to 16.35 MPa at the depth of the BSR. This makes hardly any change to the geothermal gradient calculated for the region.

Generally, the conductivity of an n-component system is given by\(^{17}\)

![Figure 5. Nomogram for velocity versus hydrate saturation calculated using porosity at BSR for the Cascadia accretionary prism along with upper and lower bounds.](image-url)
$K = \prod_{i=1}^{n} K_i^{v_i}, \quad (12)$

where, $K_i$ and $v_i$ are the conductivity and volume per cent of the $i$th component. Based on eq. (12), the background thermal conductivity–depth function of the sediment can be derived from the porosity–depth function using the following relation \cite{17} between the porosity and the thermal conductivity $K_i$ as:

$K_i = K_m^{(1-\phi)} K_w^\phi, \quad (13)$

where $\phi$ and $(1 - \phi)$ are volume fraction of water and the matrix in the sediment respectively. $K_m$ and $K_w$ are 2.5 and 0.6 W/m/K, representing thermal conductivities of the matrix and water respectively. $K_m$ was obtained by taking best-fitting geometric mean of matrix conductivities found in several continental margins of the world \cite{17}.

The thermal conductivity of the hydrated sediment, $K_{hs}$, in the depth interval between 125 to 215 mbsf can be similarly determined as:

$K_{hs} = K_m^{(1-\phi)} K_h^{3\phi} K_w^{(1-3\phi)}, \quad (14)$

where $K_h(= 0.4$ W/m/K) represents the thermal conductivity of pure hydrates \cite{17}. The solid line (Figure 7a) displays the thermal conductivity variation with depth for the background sediment. The upper and lower bounds of the thermal conductivity are also shown in Figure 7a. Thermal conductivity of hydrated sediment (shown by dotted-dashed line in Figure 7a) is less than the background value. The average thermal conductivity of the sediment up to the BSR is calculated as 1.02 W/m/K, which multiplied by the geothermal gradient gives rise to the mean heat flow of 62.85 mW/m² for this region. The estimated value matches well with the available result (65 mW/m²) \cite{16}. The bounds of estimated thermal conductivity and heat flow are 0.95 to 1.1 W/m/K, and 58 to 67 mW/m² respectively.

The formation resistivity ($R_i$) (or the resistivity of the hydrated sediment in the present case) can be calculated using the Archie’s law \cite{10} as:

$S_w^n = \frac{R_0}{R_i}, \quad (15)$

where $S_w$, $n$ and $R_0$ are respectively, the water saturation, saturation exponent and resistivity of the formation when pores are filled with water only. $R_0$ or the background resistivity can be calculated as:

$R_0 = \frac{a R_w}{\phi^m}, \quad (16)$

where $\phi$, $R_w$ and $m$ are the porosity, water resistivity and cementation factor respectively, and $a$ is a constant. The Archie’s parameters \cite{18} are taken as $a = 1$, $m = 2.8$, $n = 1.9$. $R_w$ is calculated using the following relation \cite{18}:

Figure 6. (a) Seafloor temperature vs water depth function \cite{16}, (b) Pressure–temperature phase diagram for gas-hydrate stability field \cite{17}. Solid line is the sea water curve from the equation of state computation for artificial sea water by Englezos and Bishop (1986) and Dickens and Quiaby-Hunt (1994). BSR temperatures in DSDP–ODP drill holes are also indicated.
\[
R_w = 1/\left( 3 + \frac{T}{10} \right).
\]  

(17)

where \( T \) is the formation temperature that can be derived at different depths from the geothermal gradient. The variation of resistivity with depth and error bounds are shown in Figure 7b. The average resistivity of formation is calculated as 1.13 \( \Omega \)-m that can vary from 0.89 to 1.43 \( \Omega \)-m for this region.

Location of the study area in the Makran accretionary prism is shown in Figure 2a, where presence of hydrates has been identified by a distinct BSR at 2210 m depth (495 mbsf) on the seismic section (Figure 2b). Fine-scale velocity structure (Figure 2c) across the BSR is determined by Sain et al.\(^7\) at CDPs 4375 and 4400 respectively. The velocity–depth function at CDP 4400 is considered for calculation of physical parameters, where the seafloor is at 1715 m depth and seafloor velocity is 1.78 km/s. The velocity from 285 mbsf or 2000 m depth starts increasing and reaches a maximum value of 2.2 km/s at the BSR. Since the velocities above the hydrated sediment and below the free-gas zone in the Makran accretionary prism are widely apart, and not like those in the Cascadia margin, we find difficulty in deriving the background or reference velocity–depth function deeper than the BSR depth. For the Cascadia margin, we are able to calculate the compaction factor and hence the background porosity–depth function from the data themselves. We use the compaction factor from the published literature\(^6\) to determine the background porosity depth function using the Athy’s law.

The seafloor porosity is determined using eqs (5)–(7), in a similar way as described for the Cascadia margin. We use \( V_m = 4.50 \) km/s, \( V_a = 1.5 \) km/s, \( \rho_m = 2.76 \) g/cc, \( \rho_a = 1.05 \) g/cc for sandy sediments in the Makran accretionary prism\(^9\) and \( W = 1.2 \) is taken for unconsolidated marine sediment\(^10\). Using eq. (4), the bulk density at the seafloor is calculated as 1.77 g/cc, corresponding to the seafloor velocity of 1.78 km/s. The solid line in Figure 8a shows the seafloor porosity as 54%, associated with the seafloor velocity. Assuming errors in seafloor velocity from 1.69 to 1.87 km/s produces errors in seafloor porosity from 49.5 to 59% (Figure 8a). Using the compaction factor \( \lambda = 1.44 \) per km in eq. (8)\(^9\), we calculate the background porosity–depth function and the error bounds.
**Figure 8.** a, Nomogram of velocity vs porosity for seafloor in the Makran accretionary prism along with lower and upper bounds. b, Background porosity–depth function along with lower and upper bounds. c, Background density–depth function along with lower and upper bounds.

**Figure 9.** Nomogram of velocity vs hydrate saturation calculated using porosity at BSR in the Makran accretionary prism along with lower and upper bounds. (Figure 8b). Porosity at the depth of the BSR is determined as 38.3%, that can vary from 35 to 41.8%.

The seafloor density is already calculated as 1.77 g/cc. The background density variation with depth function (solid line in Figure 8c) is calculated using the relation between the porosity and the bulk density ($\rho_b$) as:

$$\rho_b = \rho_w \phi + \rho_m (1 - \phi).$$  

(18)
The average density for the Makran region is determined as 1.98 g/cc, which can vary from 1.9 to 2.04 g/cc if an error of ±5% is assumed for the available velocity.

Using eqs (9)–(11), a nomogram (Figure 9) of hydrate saturation versus seismic velocity is prepared, from which we determine the hydrate saturation as 13%, corresponding to the maximum velocity of 2.2 km/s at the BSR. The estimated value of maximum hydrate saturation at the BSR matches quite accurately with 10% saturation as reported earlier. Taking ±5% error in velocity estimation, the maximum hydrate saturation may vary between 12.5 and 13.5%. Assuming 0% hydrate saturation at 285 mbsf and maximum hydrate saturation at the BSR, the saturation gradient is determined as 0.057% per m.

Corresponding to the water depth of 1715 m, we get the seafloor temperature as 4°C from the water depth versus seafloor temperature curve (Figure 10) obtained from the real data. The water and submarine sediments up to the BSR are converted to an equivalent pressure of 27.23 MPa, which corresponds to the BSR temperature of 21°C (Figure 6b). The difference in temperatures at the BSR and seafloor divided by the depth of the BSR with respect to the seafloor produces a geothermal gradient of 34.34°C/km.

The assumed (±5%) error in sedimentary velocity changes pressure at the BSR from 27 to 27.5 MPa, which hardly changes the geothermal gradient estimated for the region.

Using eqs (13) and (14), and taking $K_s = 2.5 \text{ W/m/K}$, $K_r = 0.4 \text{ W/m/K}$ and $K_a = 0.6 \text{ W/m/K}$, we can calculate the variation of background thermal conductivity and the error bounds (Figure 11a). The thermal conductivity of the hydrated sediment is also shown by dotted–dashed lines (Figure 11a).

The average thermal conductivity of the sedimentary formation and hence the heat flow are calculated as 1.268 W/m/K and 43.55 mW/m² respectively. The values match well with the published results of 1.27 W/m/K and 43 mW/m² for thermal conductivity and heat flow respectively. Considering ±5% error in velocity estimation, bounds in thermal conductivity and heat flow are calculated as 1.2 W/m/K and 41 mW/m², and 1.34 W/m/K and 46 mW/m² respectively.

The resistivity–depth function (Figure 11b) is derived from Archie’s equation using $a = 0.9$, $m = 2.7$, $n = 1.9$, in the same way as described for the Cascadia margin. Average resistivity of the sedimentary formation in the
Makran region is calculated as 1.96 Ω-m, which can vary from 1.54 to 2.5 Ω-m.

Two case-studies in completely different tectonic/geological environments are presented here to demonstrate how we can determine various physical parameters like porosity, density, saturation, thermal conductivity and resistivity of sedimentary formation from seismic data and the BSR. Since several relations exist among velocity, porosity and saturation, use of a particular equation may be associated with some errors in calculating hydrate saturation and several other physical parameters. As we use the velocity–depth function derived by the sophisticated waveform inversion, the velocity function can be considered quite accurate. Therefore, we have considered only ±5% error in estimating velocity function and accordingly, we get the error bounds for estimation of other physical parameters.

Presence of gas-hydrates in both the regions shows a decrease in thermal conductivity and increase in resistivity compared to the background trend. We find good agreement between heat flow derived from the BSR and the value reported earlier in the study area. This shows the strength of the approach to determine the heat flow value in a region, knowledge of which helps understand the processes involved in the earth’s dynamism. A good match between the parameters calculated from the seismic data and those measured in the nearby area indicates the suitability of the approach in determining various physical parameters without any probe data and in inaccessible areas with less expenditure. It is to be stated that the derivation of these parameters suffers from propagation errors.


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