On the use of satellite altimetry to infer the earthquake rupture characteristics: application to the 2004 Sumatra event

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SUMMARY

Standard data and methods, such as the inversion of seismic and GPS data, have been used extensively to infer the details of the 2004 December 26 earthquake. The unprecedented large size of this event gave the opportunity to modern altimeters to provide the first clear records of a tsunami in deep ocean, therefore allowing us to study the rupture history from an independent perspective. We invert the Jason-1 and Topex–Poseidon altimetry records, considering the new constraints available on the geometry of the fault plane, and taking them into account in a 3-D rupture model. The data are corrected for the non-negligible effect of satellite motion during measurements. Our results show that the rupture propagated over the 1500 km of subduction zone initially identified by the aftershock distribution, with a magnitude of $M_w = 9.1$. Our solution compares well with the latitudinal distribution of slip inferred from other data sets, with a maximum of energy release north of Sumatra, and two other slip patches near the Nicobar and Andaman islands. Based on waveform comparison, we assert that the shallow portion of the megathrust offshore Banda Aceh had slip amplitudes of more than 20 m. Also, we find that significant amounts of slip (about 10 m) concentrated below the Andaman islands and did not propagate on the shallow portion of the interface. Although synthetic tests tend to show less resolution in the northern part of the rupture, this solution is compatible with the near-field data (GPS, coral heads and imagery), and would allow one to explain the apparent paradox between the large local displacements and the moderate tsunami observed locally. Finally, we demonstrate the rapidly dominating effect of propagation and slip distribution over the rupture velocity, and how it precludes the direct estimate of this latter parameter.

Key words: Tsunamis; Earthquake source observations; Subduction zone processes; Indian Ocean.

INTRODUCTION

We recall that the Sumatra–Andaman event has the longest rupture extent ever identified, about 1500 km long, and one of the largest magnitude recorded, somewhere between $M_w = 9.1$ and 9.3 (Stein & Okal 2005; Chlieh et al. 2007). Since the subduction zone borders the western coast of the island, the earthquake caused substantial damage in northern Sumatra (Borrero et al. 2006). Indeed, most of the devastation was caused by the associated tsunami which killed over 250 000 people throughout the Indian Ocean. Above its size, this event is relevant to the scientific community as it was recorded by numerous, diverse and quality instruments which yield various solutions, and therefore, will help stress the limits of standard methods and data, and help evaluate the potential of new ones (e.g. Guilbert et al. 2005; Occhipinti et al. 2006; Smet et al. 2007).

Until recently, tsunami studies mainly rely on tide gauge measurements, witness reports or post-event field surveys. All these type of records include non-linear coastal amplifications which preclude straightforward studies of the source and of the propagation effects. To overcome this limitation, Satake (1987) proposed to invert only for the first strong oscillations of the marigrams for which linear theory still holds. Since then, this approach has been used by many authors, and turned out to be of great value, especially in the case of historical events solely recorded on tide gauges (e.g. Hirata et al. 2003a).

However, tide gauges are often characterized by poor amplitude and frequency responses, and located in complex coastal configurations such as harbours (e.g. Rabinovich et al. 2006)). Indeed, the best way to get rid of the non-linear effects is to record the tsunami wavefield before it has time to interact with the coast. There are instances of such signals through the deployment of bottom-pressure recorders (e.g. Ritsma et al. 1995; Hirata et al. 2003b) or DART buoys (Gonzalez et al. 1998).

At the turn of the century, satellite altimeters also had demonstrated their ability to identify tsunami: Okal et al. (1999) identified...
on satellite ERS-1 a sea surface height anomaly related to the 1992 Nicaragua tsunami-earthquake. The maximum amplitude of this signal was only 8 cm, hardly more than the background noise and the resolution of the instrument, and remained the only one until the 2004 December 26.

For the first time, the Indian Ocean tsunami provided high resolution satellite altimetry records of a tsunami in open ocean (Ablain et al. 2006): Topex–Poseidon (T–P), and its now successor Jason-1, happened to be flying over the bay of Bengal only 2 hr after the tsunami was triggered. Both altimeters had a SW–NE trajectory, and recorded a maximum perturbation of almost 70 cm (Fig. 1). One hour later, it was the turn of Envisat to sample the tsunami wavefield (pass 352), this time with a descending orbit. Later on, the GFO satellite captured twice the tsunami wavefield that was still invading the Indian Ocean (passes 208 and 210), but already 7 and 9 hr after the earthquake.

In this paper, we invert for the details of the earthquake rupture history using altimetry records. To this aim, we will take into account the satellite motion, the recent improvements on the 3-D structure of the subduction interface, as well as the effects of rupture velocity. Also, in our discussion, we will focus on the still debated issues of the rupture velocity, and slip distribution towards the end of the rupture.

This work is based on the Jason and T–P records, and will not consider the Envisat or GFO tracks, which have a smaller signal to noise ratio, and are more affected by non-linear coastal reflections given their time delay after the tsunami initiation.

**1 DATA AND METHOD**

### 1.1 Inversion of tsunami data

Tsunami waves can be usefully, and efficiently modelled by a simplified form of the Navier–Stokes equations, namely, the non-linear shallow water wave equations (Stoker 1955). The very small amplitudes of tsunamis in open oceans compared to water depth—less than 70 cm in the case of the Indian Ocean tsunami, while the water depth often exceeds 1000 m—allows one to neglect the contribution of the non-linear terms of the shallow-water approximation (for a comprehensive discussion, please refer to Liu et al. 1991; Synolakis 2002), and to reduce the problem to the following linear system of equations:

\[
\frac{\partial V}{\partial t} = -g \nabla \eta
\]

\[
\frac{\partial \eta}{\partial t} = -\nabla \cdot (dV),
\]

where \(V\) is the depth-averaged horizontal velocity vector, \(g\) the gravitational acceleration, \(\eta\) the tsunami amplitude and \(h\) the water depth.

We solve these equations on spherical coordinates by means of a finite-difference method, centred in time, and using an upwind scheme in space (Heinrich et al. 1998; Hébert et al. 2001). As we are only dealing with propagation in deep ocean, a coarse 2° bathymetric grid, extracted from the ETOPO2 (National Geophysical Data Center 2001) global model, will have a sufficient precision. To comply with the Courant–Friedrichs–Lewy stability condition, we impose a time step of 4 s.

In the case of tsunami generated by earthquakes, the linear theory of elastic dislocation provides a direct relationship between the slip on the fault, and the sea bottom deformation (Okada 1992). Therefore, if we assume the sea bottom deformation to be entirely and instantaneously transmitted to the above water column, the tsunami wavefield computed with eqs (1) and (2) is linearly related to the slip distribution on the fault.

Consequently, it is possible to invert for the contribution of the different parts of the fault to the satellite altimetry records through a fault plane discretization. Here, the Green’s functions correspond to tracks of the tsunami wavefield sampled at different time intervals so as to take into account the satellite motion. The Green’s functions are computed for 1 m of slip on each subfault, and their relative contribution to the altimetric signal is inverted using a least-square method: we choose the non-negative implementation (NNLS) of Lawson & Hanson (1974) which constrains the exploration to realistic solutions where slip can only be positive; this condition was initially developed for seismological inversions of the source (Kikuchi & Kanamori 1982), and is now also used in standard in tsunami studies.

### 1.2 The model parameters

To image the 1500 km of potential slip area covered by the aftershocks (Engdahl et al. 2007), and to model the effective rupture velocity, we discretized the subduction interface into 38 subelements; these elements of equal surface are evenly distributed on the upper and lower parts of the fault to identify variations of slip with depth. Furthermore, to reduce slip amplitude uncertainties related to non-contiguous, rotated, rectangular subfaults, we turned to quasi-rectangular elementary surfaces designed to fit the downdip and along strike curvatures. The deformation associated to all these
Figure 2. Green’s functions (GF) for the shallow (left-hand column) and deep subfaults (right-hand column), computed for a rupture velocity of 2 km s\(^{-1}\), and aligned with the Jason-1 altimetry profile. The black waveforms take into account the satellite motion, while the grey waveforms are instantaneous snapshots of the sea surface deformation, taken 1:55:20 h after the initiation of the rupture. The vertical line cutting through the leading peak of Jason-1’s signal allows one to identify the necessary contribution of shallow subfault number 5 to fit the altimetry data. The amplitude of the Green’s functions was multiplied in order to scale with the Jason-1 record.

The end of the locked zone as most aftershocks cluster in this part. Indeed, this later statement can even be extended over the whole rupture length according to the relocated aftershock catalogue of Engdahl et al. (2007). A last refinement comes from GPS records (Subarya et al. 2006), and satellite images (Meltzner et al. 2006; Tobita et al. 2006; Smet et al. 2007) which bound the projection of the line of no vertical change (also called pivot line or hinge line) on the surface. In particular, we have tried to reproduce its curvature towards east below the Andaman islands, and its location west of the subsided Nicobar islands.

To take all the aforementioned indications into account, we have adjusted our fault model to the following parameters: the upper part of our fault plane is taken parallel to the trench, and starts 4 km below the sea-bottom. The shallow subfaults have a dip increasing from 9° in the south (subfaults 1–6, Fig. 3), to 13° in the north (faults 10–19), while all the deep subfaults have a dip of 17°. We imposed a purely reverse fault motion up to the Andaman islands. Further north, GPS vectors (Tsai et al. 2005; Chlieh et al. 2007) suggested a rake increase to as much as 140°, somehow aligning with the azimuth of the Indian plate subduction: this variation was introduced in the model by imposing a fixed slip azimuth of N247° to all the subfaults north of latitude N7° (subfaults 7–19 and 26–38).

This correction is also consistent with a rotation of the P-axis of the main thrust aftershocks in the last part of the rupture zone (Engdahl et al. 2007).
1.3 The altimetric records of the tsunami
The Jason-1 and T–P tracks are very close in time (less than 7 min) and space (about 130 km), they have similar average height precisions (3 and 5 cm, respectively), and both sample the surface of the ocean every second. Unfortunately, it is not possible to take full advantage of the T–P record which suffered from multiple recorder anomalies (Abblain et al. 2006) (the satellite was subsequently retired in 2006 January).

In addition to the tsunami perturbation, the raw altimetric signals include perturbations brought forth by currents and eddies (Fu & Cazenave 2001). To filter these unwanted contributions, we used the correction proposed by Abblain et al. (2006) who adapted the ocean variability mapping technique of Le Le Traon et al. (1998) to this specific case. This method is more dependable than subtracting a preceding track as it integrates information provided by multiple satellite altimeters over longer and closer periods of time.

The correction of the altimetric signals for the time of the earthquake initiation can have a strong impact on the quality of the inversion as there are only a few seconds time-shifts between the Green’s functions (Fig. 2). For this reason, the Jason-1 and T–P tracks are corrected using the best time accuracy computed by the USGS (00:58:53.49 GMT for the tsunami tracks are corrected using the best time accuracy computed by the USGS (00:58:53.49 GMT ±0.14).

Finally, it is of basic importance to consider that the Jason-1 and T–P satellites started to intersect the tsunami wavefield only 2 hr after its initiation, and because of their limited ground speed (around 7 km s⁻¹), it took them more than 8 min to go across the already widely extended wavefield. In other words, the altimetry signals are not exact snapshots of the tsunami wavefield, as was considered in previous studies, but rather a time-evolving samplings of the wavefield along a fixed track. Thus, in order to compute each Green’s functions, we extracted 20 segments from different instantaneous profiles of the sea surface (the Green’s function in the case of an instantaneous satellite recording) that would match the constant progression of both the tsunami and the satellites (Fig. 2).

2 SYNTHETIC TESTS AND RESULTS

2.1 Synthetic tests
Synthetic tests, where you attempt to invert the altimetry profiles computed from a predefined source, allow one to check the robustness of inversions, to identify the level of uncertainties, and possible trade-off among the inverted parameters. Seismological studies have demonstrated that strong trade-off may arise from the inversion of both static (amplitude), and kinematic parameters (rise-time, rupture velocity) (Konca et al. 2007); the inversion of altimetry data is somewhat a different problem but such effects may also arise. Hence, we performed synthetic tests, first looking at the slip amplitude resolution, and then at the effect of rupture velocity.

For any given slip distribution, the inversion scheme was systematically able to recover the original patterns and amplitudes, and even with some strong noise added to the Green’s functions, the solutions still match very well. This can be explained by the strong time delay between all the Green’s functions, both along strike and at depth, but also by their different frequency content (Fig. 2). Several other factors contribute to the stability of these inversions such as the linear formulation of the problem, the positivity constraint on slip, and the absence of timing errors (rupture velocity is known). Synthetic tests have also been performed using Green’s functions generated using non-linear shallow water equations, and led to comparable results. This later result was expected as both types of Green’s functions are extremely similar, and support the assumption of a linear behaviour for the tsunami propagation.

Then, we proceeded with some tests to check the impact of rupture velocity errors of ±0.5 km s⁻¹ on a reference model with a prescribed rupture velocity of 2 km s⁻¹ (Fig. 3). Several other tests have been conducted, using different initial slip distributions and rupture velocities, but as all solutions exhibited the same effects, only one representative test is presented here. For both rupture velocities, 1.5 and 2.5 km s⁻¹, the results are very different between the first and the second halves of the rupture. The first 600 km are always well resolved, with slip distributions and moment release similar to the reference model. The slip can increase locally (over one subfault), but then it is usually compensated by a lower slip on one of its neighbouring along strike subfaults. Therefore, we can pretend to have a good resolution at the length scale of two subfaults along the strike (about 150 km), and at the length scale of one subfault along the dip (about 70 km).

For the second, northern part of the rupture, the results are not as good as the inversion tends to smear the energy of the shallow segments towards the deeper ones. As a consequence of the geometry of the problem and of rupture propagation, the timing error between Green’s functions is much more pronounced in the northern half of rupture (see the increase in delay between Green’s functions on Fig. 2). Thus, with an incorrect rupture velocity, the inversion scheme can more easily reduce the misfit by fitting the low frequency component, that is using low frequency-deep Green’s functions rather than using the high frequency shallower ones. Consequently, the altimetry profiles produced by the inversion do not adjust the small and high frequency peaks recorded by the satellites towards the end of the rupture, but manage to fit the low frequency oscillations quite well, and provide a correct value of the seismic moment (ΔM0 = 0.32 × 10²² N m, equivalent to a moment magnitude error of 0.02).

2.2 Results of the inversions
Fig. 4 presents the results of the joint inversion of Jason-1 and T–P tracks performed for three different rupture velocities, 1, 2 and 3 km s⁻¹, spanning the range of values proposed so far (Ammon et al. 2005; Guilbert et al. 2005; Fuji & Satake 2007). The computed altimetry profiles closely match all the details of the corrected Jason-1 and T–P records.

As expected from the previous synthetic tests, the slip distributions inverted (Fig. 4) are very consistent whatever the rupture velocity. First, they are strongly dominated by a very localized slip patch offshore Banda Aceh and 300 km north of the hypocentre, with up to 40 m of displacement at depth (Fig. 4). These extreme values are derived for the slowest rupture velocities (1 and 2 km s⁻¹) while slip seems more widespread towards the hypocentre for 3 km s⁻¹. The slip amplitude of shallow segment number 5 varies between 26 and 32 m. At the latitude of the Nicobar islands, between 7 and 9°N, there is a second persistent zone of substantial slip with average amplitudes of 8 m. Third, there is a widespread zone of strong slip below the Andaman islands, whose amplitude fluctuates between 8 and 22 m depending on rupture velocity. Finally, we can point out that the slip distributions systematically stop at the northern tip of the Andaman islands, and do not extend further north, beyond the aftershock area.

All these features are very consistent, at least in terms of inversion, as the corresponding synthetic altimetric signals are almost
overlapping and explain equally well the records (Fig. 5b): the normalized root mean square (rms) are very low (below 0.4), and characterized by a very small deviation ($\Delta \text{rms} = 0.02$). The seismic moment is also very stable around $5.4 \times 10^{22}$ N m, which corresponds to $M_w = 9.1$. These values are computed assuming a rigidity with a standard value of $4 \times 10^{10}$ N m$^2$.

3 SUMMARY AND DISCUSSION

3.1 Beginning of rupture and high energy release

The inversion distinctly locates the maximum slip patches in the area off Banda Aceh (subfaults 5 and 24, Fig. 4). This result is coherent with the several field surveys conducted in the area (Tsuji et al. 2005; Jaffe et al. 2006; Lavigne et al. 2007), with the seismological and geodetic inversions (Ammon et al. 2005; Vigny et al. 2005; Chlieh et al. 2007), but also with the recording of the highest amplitudes at the onset of the satellite records as this section of the subduction has the shortest propagation path to the satellite tracks. The latitudinal position of this maximum is also consistent with what other authors (Hirata et al. 2006; Fujii & Satake 2007; Piatanesi & Lorito 2007) have obtained from tsunami data.

However, over these last solutions, the downdip discretization of our model provides valuable additional refinements on the slip model: as the surface deformations are more filtered with increasing depth of the fault plane, the Green’s functions corresponding to shallow subfaults contain more high frequencies and have a sharper wave front than their deeper counterparts. Hence, simply comparing these Green’s functions reveals that the shape of the two leading, narrow and high amplitude peaks of the Jason-1 record can only be adjusted by slip on the shallow segments (Fig. 2). This assertion is reinforced by the timing of the deep segments whose Green’s functions arrive too late to model the first peak of Jason-1 altimetry profile, but also the dense concentration of aftershocks near the trench (Fig. 1) on this portion of the subduction megathrust (Geist et al. 2006). Synthetic tests presented in the previous sections have also shown that the solutions obtained for this part of the rupture are very robust.

These results also highlight the complementary nature of tsunami data with other types of measurements: Vigny et al. (2005), Chlieh et al. (2007) solely inferred slip at depth likely because this fault segment was only constrained by distant GPS vectors—located more than 200 km away from the trench—while it is well known that teleseismic data poorly constrain the vertical distribution of slip (Christensen & Ruff 1985; Wagner & Langston 1989).

The amplitudes obtained (Fig. 4), 26–40 m, are on average slightly above the values of 25–30 m computed from tsunami data by Hirata et al. (2006), Piatanesi & Lorito (2007) and Fujii & Satake (2007). This discrepancy is most likely related to the finer discretization of the model, and the concentration of slip on a few small subfaults, as the total moment is similar. In any case, all the tsunami derived amplitudes, including other studies, are far above the maximum 15 m inferred by the seismological model of Ammon et al. (2005), and closer to the values inferred by Vigny et al. (2005) or Chlieh et al. (2007) from GPS data.
The main difference among the three solutions presented in this study corresponds to the first 300 km of rupture, preceding subfaults number 5 and 24. For all three rupture velocities tested, the slip distribution is physically plausible with a minimum of 8 m of slip on the epicentral subfault. However, considering the fit to the altimetry waveforms, the variations are too subtle to discriminate between the models. Only the field measurements of uplifted and subsidised coral heads, plus one GPS vector right north of Simeulue island, provide a strong constrain on the amount of slip at the hypocentre: Subarya et al. (2006) and Chlieh et al. (2007) gave an estimate of 10 m slip, close to the values inferred from inversion with rupture velocities of 1 and 2 km s\(^{-1}\) (Fig. 4). Ammon et al. (2005) found amplitudes of the same order based on the inversion of seismological data.

### 3.2 Slip distribution below the Andaman islands

Results of the inversion, for all rupture velocity tested, require a significant amount of slip at the latitude of the Andaman islands. This issue, which has long been raised, is now well established from satellite images, hyperspectral analysis and the re-occupation of the GPS benchmarks (Tobita et al. 2006; Chlieh et al. 2007; Smet et al. 2007). Even so, all these data remain limited to the surroundings of the islands, and hardly constrain the distribution of slip up to the trench given their proximity to the pivot line (Subarya et al. 2006).

Our fault discretization fosters the interpretation of a slip confined to the deep segments defined below the Andaman islands. Although the synthetic tests suggest that this part of the fault plane might not be well resolved, all inversions done at intermediate rupture velocities between 1 and 3 km s\(^{-1}\) (with increments of 0.2) require a complete lack of slip on the shallow segments. Furthermore, there is this external argument of the almost absence of seismicity offshore the Andaman islands, while many events are localized along the western coast, or below these islands (Fig. 1). This lack of events can be the consequence of the large accumulation of unconsolidated sediments deposited by the Ayeyarwady and Ganges rivers, and interpreted as a marker of the upper limit of the coseismic rupture. Interestingly, Konca et al. (2007) have reached a similar conclusion for the 2005 March 28 Nias earthquake (\(M_w = 8.6\)); this event followed the December rupture, and occurred on the adjacent south segment, a region of comparable geometry. Thanks to an extensive set of uplift-subidence field measurements, and GPS benchmarks located right above the area of rupture, the aforementioned authors were able to precisely locate the main slip patches below the islands. Also, as in the case of the Andaman rupture, the aftershocks densely concentrated along the western coast of the islands, suggesting a similar mechanical response to the March event (Hsu et al. 2006).

### 3.3 Temporal resolution

In the case of very long ruptures, it is now well established that the slip heterogeneity on the fault can have an influence on the far field amplitudes of a tsunami (Hébert et al. 2007). On the contrary, the kinematic effects have always been discarded given that tsunami propagate slower by approximately an order of magnitude. The rupture velocity is thus commonly assumed to be infinite while its influence on real data has never been demonstrated (Geist 1999).

The altimetric records combined with the exceptional rupture length and duration of the Sumatra earthquake—\(T\) wave (Guilbert et al. 2005) and high frequency seismograms analysis (Lomax 2005; Ni et al. 2005) inferred a duration of more than 500 s—should therefore be a good opportunity to investigate this parameter. Moreover, the contribution of a potential slow slip component is still unresolved: Stein & Okal (2005) concluded, based on the analysis of normal-modes, that the rupture included a slow component, which would have ‘probably occurred over the northern part […] of the rupture zone’. Later, Singh et al. (2006) based on the re-evaluation of the Port-Blair tide gauge record (Andaman islands), suggested that a significant part of the slip occurred between the passage of the initial rupture front and 35 min.

In spite of all these unresolved issues, the results of the inversion presented above, and especially the synthetic tests, tend to bear out the inability to resolve the rupture velocity using the altimetry data alone, at least not in the configuration of the Sumatra event. Fuji & Satake (2007) also pointed out a lack of resolution, and could hardly gain in sensitivity with the addition of marigrams in the inversion process. Inversions performed at eleven different rupture velocities between 1 and 3 km s\(^{-1}\) did not reveal any significant change in the misfit, although solutions between 2.6 and 3 km s\(^{-1}\) had slightly higher rms.

In order to comprehend this lack of resolution, we can stress that at least one effect has screened the influence of rupture velocity: the Jason-1 and T–P satellites propagated north with similar directions and velocities (\(\geq 7\) km s\(^{-1}\)) to that of the rupture, therefore reducing the apparent rupture velocity on the recorded profiles. Regarding the kinematic effects, it is also more difficult to tight the timing of the rupture as the altimetry profiles were recorded after 2 hr of propagation: the time delay reduced the ability to identify distinct phases of the tsunami wave pattern which have been generated along different parts of the fault. Hence, the lack of resolution for the rupture velocity, which can also be interpreted as a trade-off with the slip distribution, merely expresses the rapidly dominating effect of tsunami propagation (geometrical spreading).

In this study, we chose not to address the issue of a possible slow slip component, and leave this question for future studies. However, it is worth noting that we were able to match the altimetry records in great detail without the need of rupture rise times of more than 10 min. This indirect evidence adds to the analysis of the continuous GPS stations settled in Malaysia and Myanmar which did not show any sign of slow motion (Vigny et al. 2005), and to the detailed inspection of the long-period seismograms which led to a similar conclusion (Velasco et al. 2006).

### 4 CONCLUSIONS AND PERSPECTIVES

Through the use of only the Jason-1 and T–P altimetry profiles of the tsunami, we were able to infer a consistent and coherent description of the earthquake slip distribution. The unprecedented 1500 km of rupture, extending from the Simeulue island to the north of the Andaman archipelago as initially suggested by the aftershock distribution (Engdahl et al. 2007), are supported by our inversions. In agreement with most models published so far, we inferred a maximum energy release between 3.5° and 5.5° N. Also, the modelling of the tsunami data makes it mandatory to have a significant amount of slip on the shallow segments of this area; there, some active splays faults were recently imaged by OBS and deep seismic campaigns (Klingelhoefer et al. 2006; Sibuet et al. 2006), but it is still not clear whether they significantly contributed to the tsunami generation.

Although difficult to model entirely using dislocation models (Chlieh et al. 2007), the geodetic measurements remain compatible with the strong slip amplitudes of 10 m or more inverted for the northern part of the rupture. Moreover, our solution is compatible with the weak amplitudes of the tsunami as the earthquake...
deformation, confined to areas of shallow waters (i.e. around the Andaman islands), would have displaced limited amounts of water. Even though the along strike variations of seismic moment are supposed to be well constrained, the depth distribution still needs confirmation as the synthetic tests indicate a lower resolution in this part of the rupture. In any case, the concentration of slip below the Andamans would reveal striking similarities with the slip pattern of the following 2005 March 28 Nias earthquake (\(M_w = 8.6\)). One straightforward interpretation of this finding is that rupture mechanisms identified for the well imaged Nias rupture (Hsu et al. 2006) can be invoked for the Andaman region, and possibly expanded to the whole Sumatra subduction zone.

In spite of the precise and fine discretization used, the inversion of Jason-1 and T-P profiles did not display much sensitivity on rupture velocity to constrain its value, stressing the rapidly dominating effect of propagation and slip distribution over the kinematic effects. Yet, the slight increase of the rms for rupture velocities above 2.6 km s\(^{-1}\), combined with a slip distribution less compatible with the geodetic estimates for the beginning of the rupture, slightly reduce the likelihood of a high rupture velocity.

Beyond the analysis and results presented in this study, many aspects of the open ocean records remain to be explored. Yet, more tsunami records in open ocean are expected with the launch of new altimetry satellites (Jason-2 should start to operate in 2008), and the ongoing deployment of DART buoys and bottom-pressure sensors. These new generation data call for more in-depth analysis of their content and limitations given their great potential to help understand the earthquake rupture and tsunami propagation.

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